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Salinity Thresholds, Lake Size, and History: A Critique of the NAS and CORI Reports on Mono Lake

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Abstract.—These two reports usefully summarize and evaluate large amounts of information. They both suffer from and perpetuate, however, a deficient conceptual framework for analyzing changes in the Mono Lake ecosystem. Specifically, their assessments of the present and future state of this lake 1) use language implying that salinity-induced changes in the biota will begin to occur only at certain critical salinity thresholds, 2) neglect the significance of lake size as a determinant of bird food supplies, and 3) lack historical perspective in failing to consider what changes in the lake ecosystem may have been caused by historical changes in the salinity and size of the lake. The desirability of developing explicit models for the system is emphasized. Especially needed are models for: 1) the influence of salinity and lake size on the abundance of brine flies and brine shrimp, and 2) the influence of the abundance of these invertebrates on the bird populations that use the lake. To illustrate the heuristic value of such models, Rawson's models relating productivity and standing crop to lake mean depth are applied to Mono Lake. The results suggest some unusual consequences of the lake's particular morphometry, especially for lake level changes between 6370 ft and 6380 ft.

In the last few years, two major reports on Mono Lake have been published. *The Mono Basin Ecosystem* (Patten et al. 1987) was prepared at the request of the U.S. Congress by a committee appointed by the National Research Council. It was published by the National Academy of Sciences in the summer of 1987 and will be referred to hereafter simply as the NAS report. The committee's mandate was to inventory and describe the Mono Lake ecosystem, its hydrology and its populations, to review historic changes in the system, to determine "the critical water level needed to support current wildlife populations," and to predict how the system would respond to continued diversions of water from the basin.

The Future of Mono Lake (Botkin et al. 1988) was prepared at the request of the California State Legislature by a committee assembled by the Community and Organization Research Institute (CORI) at the University of California, Santa Barbara. It was published in the spring of 1988 by the University of California Water Resources Center and is referred to hereafter as the CORI report. This committee's primary mandate was to "evaluate the effects of declining lake levels, increasing salinity and other limnological changes of Mono Lake upon" the various populations living in and near the lake, including human populations.

¹ The author is a consultant to the Los Angeles Department of Water and Power. The opinions expressed in this article are his own.

The two studies overlapped greatly in their objectives, utilized mostly the same sources of information, and not unexpectedly have some similarities. They represent reasonable summaries of existing information and opinion on Mono Lake, and in that regard are useful. No other recent synthetic treatments on the ecology of Mono Lake exist.

Both reports, however, are based on and present to the reader deficient conceptual frameworks. The principal deficiencies are three and all relate to the central issue of how water diversions from the Mono Lake basin affect the organisms that live in or on Mono Lake.

First, in discussing salinity effects, both reports use language and graphical presentations that imply *thresholds*. They suggest that changes in the biota will begin to occur at certain critical salinity levels rather than occur continuously with salinity change. The latter is more likely to be the case.

Second, the importance of *lake size* is largely ignored. Obviously this will be a major factor in determining, for example, the total amount of food available to waterbird populations.

Third, the reports give a very incomplete framework for viewing *historical changes* in the Mono Lake ecosystem. Since biological data on the lake prior to the mid 1970s are very few, concrete conclusions about the past status of Mono Lake populations are not to be expected. But the reports do not even mention, for example, the likelihood that changes in the size and salinity of Mono Lake since 1941 have had effects on the lake's biota.

By neglecting the above, the NAS and CORI reports failed to provide a clear and coherent conceptual framework for management decisions, for legal decisions, and for the planning of future scientific research. The reports are less useful than they might have been and, especially for nonscientists, can be misleading.

Let me now document the above charges and in doing so suggest a broader conceptual framework for analyzing some of the anticipated effects of water diversions on the Mono Lake biota.

Certain issues will not be addressed here, such as nesting of the California Gull, substrate relations of brineflies, stream incision, tufa towers, and alkali dust problems. The focus will be on the trophic relationships among Mono Lake populations.

The benighted state of certain professions in the United States has resulted in use of the English system of measurement in most of the documents on Mono Lake. I follow this custom in my text. In my tables, however, I present values for lake dimensions in both English and metric units and values for productivity and standing crop only in metric units.

Salinity Thresholds?

There is every reason to believe that, both individually and collectively, the scientists preparing these reports were aware that the abundances or productivities of populations can be expected to change continuously as salinity is gradually increased over any given range. Changes that begin to occur only at particular salinity thresholds simply neither have been documented experimentally for any species nor are to be expected on physiological grounds. Sure, there will always be a point where a slight further increase in salinity from S_1 to S_2 will cause extinction of a population (Fig. 1A, species X_3). But under salinity S_1 the abun-

dance or productivity of the population usually would already have been much lower than under optimal salinity conditions (e.g., a salinity $\ll S_1$). In the context of conservation, the battle has already been lost if salinities have increased to close to the extinction level for the species for which protection is sought. It is the other end of the curve that should occupy our attention. What kind of reduction in abundance is acceptable or tolerable? 0%? 5%? 10%? 20%? 40%? And at what increased salinity level will such a reduction be observed?

Regardless of the committees' understanding of the above points, the language of their reports fails to convey that understanding to the reader. Threshold salinity values are stated for every species or group.

Beyond those threshold values, effects are variously said to be "significant," "critical," "very slight," "severe," etc. In no case are the meanings of those terms defined either in terms of the magnitude of the expected effect, its importance to the rest of the ecosystem, or its societal acceptability. And below those threshold values, the usual implication is that salinity variations are without effect.

Here are some examples from the NAS report:

1. "[algal] productivity is likely to decrease gradually at salinities above about 100 g/l . . ." (p. 4).
2. "Brine shrimp are expected to gradually decrease in abundance if salinities exceed 120 g/l . . . reduction . . . would be large at salinities greater than 130 g/l . . ." (pp. 4-5).
3. "The decrease in availability of brine shrimp for food would begin to affect those birds relying on them . . . at a salinity of 120 g/l . . ." (p. 5).
4. "salinities above 133 g/l significantly depress [brine shrimp survival and growth]" (p. 89).
5. "algae . . . [and] brine shrimp and brine fly populations flourish [at salinities of] <50-89 g/l" (pp. 188-189).
6. "At salinities around 120 g/l, [the algae, brine shrimp, and brine fly] populations . . . would begin to show negative responses . . ." (p. 206).

And some examples from the CORI report:

7. "food for grebes and phalaropes declines significantly [at salinities >120 g/l]" (p. 9).
8. "Algal food supply for brine fly declines rapidly [at salinities >130 g/l]" (p. 10).
9. "a decline in status of lake ecosystem begins to occur . . . where salinity reaches a value of 97 g/l . . . the growth of brine shrimp and brine flies, and the abundance of certain algae, begins to decline" (p. 12; but see p. 10 where such decline is predicted to occur "before this").
10. "brine shrimp reproduction begins to decline at a salinity of 120 g/l" (p. 24).

Figure 1 contrasts the types of salinity effects models implicit in the NAS and CORI reports with the types of models that would be expected on the basis of physiological and ecological principles. Experimental data are not available, either for Mono Lake species or for other saline lake species, for rigorous empirical tests of which type of model most closely approximates reality. One would need *precise*

estimates of population size (or productivity) at each of *many* salinities over a *wide* salinity range.

A further implication of the reports is that the threshold salinities are all higher than any salinities that have been experienced by the lake during historic times (quotes 1–10, above). Everything has “flourished” up to now or at least up to 89 g/l (quote 5). Salinity effects, if they are to occur at all, lie in the future. Such a notion is biologically indefensible. Every population in the lake surely has been affected by the rise in salinity from 51 g/l in 1941 to the 87–98 g/l experienced during the 1980s. Naturally we have little direct evidence as to what those effects have been.

The notion of “plateau and threshold” response curves and the notion that the thresholds all lie at higher than current salinities may subtly convey another idea to readers of the NAS and CORI reports: the idea that if, with increasing salinity, we are indeed approaching the edge of a “plateau,” we should proceed extremely cautiously. Such a view suggests that we should consider with alarm the prospect of even small, temporary further increases in salinity, for the edge of the plateau may be a “cliff” (species X_1) or an extremely steep “slope” (species X_3 , Fig. 1). The past may give no hint of the future. The levelness of the plateau behind us, the historical “flourishing” of the system, may be deceptive and provide no warning of the imminence of the “crash” of a population or of the whole system. And the crash may be irreversible.

Such an argument is perfectly sound on naive theoretical grounds. It is also the most rational and societally responsible position to take when little is known about system behavior or when some of the anticipated negative consequences indeed seem likely to be irreversible.

I would argue, however, that we know enough about the Mono Lake ecosystem, about other alkaline saline lakes, and about the responses of organisms in general to salinity to be confident that such alarm is unwarranted. Population response curves such as those for species X_1 and X_3 in Fig. 1A are unknown in the biological literature. There is no reason to think that increasing salinity by, say, 10 percent over current levels would by itself cause large or irreversible changes to any Mono Lake population.

Two caveats should be offered here. I am saying that a small, temporary change is not cause for alarm; but I am not saying that a small, permanent change is necessarily societally acceptable. Also, as stated earlier, I am not discussing effects directly attributable to changes in water level, e.g., peninsularization of Negit Island, rather than to salinity *per se*.

Though the “plateau and threshold” models are implied by most of the language in the reports, one doubts that any of the biologist authors of those reports intended to advocate such models or their corollaries. All those authors most likely would agree that “smooth curve” models (Fig. 1B) are more realistic and that small (e.g., 10 percent) salinity increases are unlikely to cause population or system “crashes” of any sort. I infer this from a few rare and very general statements such as “Responses of various resources to changes in lake level will, for the most part, occur gradually over a range of levels” (NAS report, p. 7), though such statements neither refer explicitly to salinity effects nor exclude “plateau and threshold” models such as that for species X_2 (Fig. 1A).

If my suppositions are correct, the discrepancies between report language and

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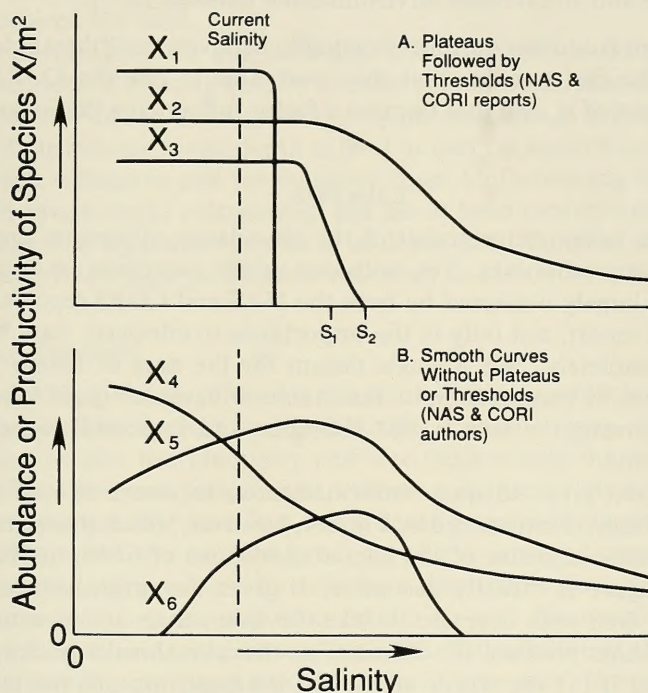


Fig. 1. Comparison of (A) the types of salinity effect models implicit in the NAS and CORI reports, and (B) those expected on the basis of physiological and ecological principles.

author opinion probably arose from failure to define terms (e.g., what is “significant” or “severe” or “critical?”) and the lack of explicit models for salinity effects on populations.

Some scientists may also argue that it was desirable to use the “threshold” terminology in the NAS and CORI reports so as to make them intelligible to lawyers, engineers, managers, politicians, and other non-scientists. The notion is without merit. Both lawyers and engineers have told me that, at least on good days, they are capable of understanding the concept of gradual, continuous change. They also have stated that if biologists, for example, state or imply that the pattern of expected biological change is likely to follow a “plateau and threshold” model, then that is exactly what they will assume represents the best collective biological judgment on the matter, in the absence of clear arguments to the contrary.

Among academics this matter might be considered only a minor one of semantics. In the larger politico-legal context of the Mono Lake controversies, however, the details on language can be very consequential. There is no better evidence of this than the “Memorandum of points” that accompanied the National Audubon Society’s (1989) recent successful petition for a writ of injunction against the Los Angeles Department of Water and Power. Among the first statements that the court found in that well-crafted document are:

“we are *now* [my emphasis] faced with the imminent and irreversible injury to Mono Lake [p. 1] . . . the lake is poised on the *threshold* [my emphasis] of substantial and irreversible environmental damage [p. 7].”

Whether or not Audubon's attorneys actually believed this “threshold” scenario is irrelevant. The fact remains that they were able to cite the CORI and NAS reports in support of it, and that became a factor influencing the judge's decision in their favor.

Lake Size

Lake size is a major determinant of the abundance of aquatic organisms in Mono Lake or any other lake. Yet, with one minor exception, its importance in this regard was largely neglected by both the NAS and CORI reports.

In the CORI report, not only is the importance overlooked, data on lake size are omitted completely. Not a single datum for the area of Mono Lake, past, present, or future, is presented in its text, table or figures. Figure 1 in the report gives a map showing the lake's 1987 shoreline and a second, higher shoreline which is unlabeled.

The NAS report gives adequate information on lake area in its Figs. 3.1 and 6.2. Slight confusion is introduced by Fig. 6.1, however, which shows, in 4 separate maps, the shoreline and size of the lake at elevations of 6340, 6360, 6380, and 6400 ft. This figure is visually deceptive. It gives the impression of a smaller decline in lake area with decrease in lake elevation than would actually be the case. Two elements produce the illusion: as the lake shrinks in dropping from 6400 ft to 6340 ft (1) the width of the shaded band around the lake remains approximately constant in width, the outer edge of the band being whittled away as the inner edge follows the receding shoreline, and (2) the scale bar representing 30,000 ft gradually increases in length by about 18 percent. The latter causes the decrease in lake area between 6400 ft and 6340 ft to *appear* to be about 27 percent when in fact the actual decrease would be about 48 percent.

The importance of lake area derives from the simple fact that the amount of solar radiation received by the lake ecosystem is directly proportional to that area. Total solar radiation received, in turn, should be a major determinant of the total primary and secondary production that takes place in the lake. Specifically, as lake area decreases the total amount of brineflies and brine shrimp potentially available to the grebes, phalaropes, gulls, and other waterbirds at Mono Lake should be expected to decrease, even if the concomitant increase in salinity has no effect.

Whether the numbers of these birds that feed at Mono Lake is currently limited by these food supplies is a complex and unanswered question (see below). But surely our main concern should be for the *numbers* of birds that utilize the lake and not their *densities*, i.e., not the numbers per unit area. The aesthetic and ecological value of Mono Lake should be a function of, *inter alia*, the numbers of birds it supports. Also, for any given species, the number of individuals that utilize the lake is a measure of the extent to which the lake contributes to the survival and health of that species as a whole. If the lake were to decrease in area by 50 percent, there would be strong cause for concern even if the density (number/km²) of birds on the lake remained unchanged.

It is not clear that the same argument is valid for the algal and invertebrate populations. A 50 percent decrease in total phytoplankters or total brine shrimp numbers might be of little concern, so long as it had no effect on the numbers of birds that utilized the lake.

The relationship between lake area and total lakewide productivity of Mono Lake's populations is complicated by another relationship, that between productivity and depth. Among lakes, productivity per unit area is widely considered to increase with decreasing mean depth at least in part because of increased nutrient fluxes between sediments and the euphotic zone. Unfortunately the exact nature of the depth-productivity relationship has never been determined for any set of lakes in a way that separates the effects of depth from those of all the lake morphometric and hydrologic parameters that tend to covary with mean depth.

A Modelling Approach

Rawson (1952, 1955) developed a simple, tentative approach to this problem that may be usefully applied to Mono Lake. It entails using mean depth as a general index of lake morphometry and size determining the mathematical relationships between mean depth and various measure of productivity *per unit* area. (I use 'productivity' here in a broad sense to include standing crop as well as productivity *sensu strictu*). The products of lake area and these productivities then give whole lake productivity estimates that reflect both the positive relationship between lake area and total insolation and the negative relationship between lake depth and nutrient cycling rates. Calculation of these whole lake estimates is actually an extension of Rawson's approach. It assumes that productivity per unit area is independent of lake area when mean depth is held constant.

Using data from 11–20 lakes in western and central Canada (that ranged from 11 m to 249 m in mean depth), Rawson used regression analysis to deduce the following three relationships:

$$FP/U = 14.71 (D^{-0.7029}) + 0.56 = \text{fish production (kg/ha/y)} \quad \text{Eq. 1}$$

$$PS/U = 3765 (D^{-1.5337}) + 8.0 = \text{plankton standing crop (kg/ha)} \quad \text{Eq. 2}$$

$$\text{and} \quad BS/U = 69.2 (D - 5)^{-0.788} = \text{benthos standing crop (kg/ha)} \quad \text{Eq. 3}$$

where D = mean depth (m).

I have used the first two of these to calculate the expected "fish" production and plankton standing crop for Mono Lake at different elevations between 6330 ft and 6430 ft. Results are expressed on both a per unit area basis and a lakewide basis in Tables 1 and 2 and Figure 2. The salinity, lake area, and mean depth corresponding to each elevation are also given, as tabulations of the values for these variables neither were presented in the NAS and CORI reports nor are available elsewhere in the open scientific literature.

Of course there are no fish in Mono Lake, but perhaps the production of (or consumption by) their trophic counterpart, the invertebrate-eating waterbirds, is a rough analogue. And of course we do not know how reliable these particular expressions are as predictors of even the relative, let alone the absolute, degrees

Table 1. Variation of the surface area, mean depth and salinity of Mono Lake as a function of the elevation of its surface. Data from tables and models in LADWP (1986, 1987). Predicted salinity values assume no net loss of salts via precipitation or other processes.

Elevation		Surface area		Mean depth		Predicted salinity		Observed salinity
(ft)	(m)	(sq. mi.)	(sq. km)	(ft)	(m)	(g/l)	(g/kg)	(g/l)
6428	1959.3	89.7	232.2	86.1	26.2	42.4	41.0	
6427	1958.9	89.3	231.2	85.6	26.1	42.9	41.5	
6426	1958.6	88.9	230.2	84.8	25.9	43.4	42.0	
6425	1958.3	88.5	229.1	84.3	25.7	43.9	42.4	
6424	1958.0	88.0	227.8	83.7	25.5	44.4	42.9	
6423	1957.7	87.7	227.1	83.0	25.3	45.0	43.4	
6422	1957.4	87.4	226.3	82.2	25.1	45.6	44.0	
6421	1957.1	87.1	225.5	81.7	24.9	46.1	44.5	
6420	1956.8	86.7	224.5	80.9	24.7	46.7	45.0	47.0
6419	1956.5	86.3	223.4	80.4	24.5	47.3	45.5	
6418	1956.2	86.1	222.9	79.5	24.2	47.9	46.1	
6417	1955.9	85.8	222.1	78.7	24.0	48.5	46.6	51.3
6416	1955.6	85.6	221.6	77.9	23.8	49.0	47.1	
6415	1955.3	85.2	220.6	77.4	23.6	49.7	47.8	
6414	1955.0	84.8	219.5	76.7	23.4	50.3	48.4	54.0
6413	1954.7	84.6	219.0	75.8	23.1	51.0	49.0	
6412	1954.4	84.4	218.5	75.0	22.9	51.7	49.6	
6411	1954.1	84.1	217.7	74.5	22.7	52.4	50.3	56.3
6410	1953.8	83.7	216.7	73.6	22.4	53.0	50.9	
6409	1953.5	83.3	215.7	73.1	22.3	53.9	51.6	
6408	1953.2	83.0	214.9	72.2	22.0	54.6	52.3	
6407	1952.9	82.7	214.1	71.5	21.8	55.4	53.0	58.1
6406	1952.5	82.4	213.3	70.8	21.6	56.1	53.6	58.6
6405	1952.2	82.0	212.3	70.2	21.4	56.9	54.4	
6404	1951.9	81.6	211.3	69.5	21.2	57.8	55.2	
6403	1951.6	81.2	210.2	68.9	21.0	58.6	56.0	60.2
6402	1951.3	80.9	209.5	68.0	20.7	59.4	56.7	
6401	1951.0	80.6	208.7	67.2	20.5	60.4	57.6	
6400	1950.7	80.0	207.1	66.8	20.4	61.2	58.3	
6399	1950.4	79.5	205.8	66.3	20.2	62.2	59.2	
6398	1950.1	79.1	204.8	65.6	20.0	63.3	60.2	
6397	1949.8	78.8	204.0	64.9	19.8	64.1	61.0	
6396	1949.5	78.4	203.0	64.1	19.6	65.3	62.0	
6395	1949.2	77.9	201.7	63.6	19.4	66.1	62.8	
6394	1948.9	77.3	200.1	63.0	19.2	66.9	63.5	
6393	1948.6	76.9	199.2	62.4	19.0	68.3	64.7	
6392	1948.3	76.5	198.1	61.7	18.8	69.3	65.6	
6391	1948.0	76.1	197.0	61.0	18.6	70.5	66.7	
6390	1947.7	75.5	195.5	60.5	18.4	71.5	67.6	
6389	1947.4	74.8	193.7	60.0	18.3	72.9	68.8	
6388	1947.1	74.4	192.6	59.4	18.1	73.9	69.7	
6387	1946.8	74.0	191.6	58.7	17.9	75.4	71.1	
6386	1946.5	73.5	190.3	58.1	17.7	76.5	72.0	
6385	1946.1	72.6	188.0	57.7	17.6	78.0	73.4	
6384	1945.8	71.4	184.9	57.8	17.6	79.6	74.9	
6383	1945.5	70.9	183.6	57.1	17.4	80.8	75.8	
6382	1945.2	70.4	182.3	56.6	17.2	82.1	77.0	
6381	1944.9	69.8	180.7	56.1	17.1	83.7	78.4	
6380	1944.6	68.7	177.9	55.9	17.1	85.2	79.7	89.3
6379	1944.3	66.8	172.9	56.4	17.2	86.7	81.0	

Table 1. Continued.

Elevation		Surface area		Mean depth		Predicted salinity		Observed salinity
(ft)	(m)	(sq. mi.)	(sq. km)	(ft)	(m)	(g/l)	(g/kg)	(g/l)
6378	1944.0	66.2	171.4	56.0	17.1	88.8	82.9	86.8
6377	1943.7	65.6	169.8	55.8	17.0	89.9	83.8	91.6
6376	1943.4	64.8	167.8	55.3	16.8	90.8	84.5	89.3
6375	1943.1	63.3	163.9	55.5	16.9	92.7	86.3	93.4
6374	1942.8	61.1	158.2	56.5	17.2	94.7	87.9	
6373	1942.5	59.3	153.5	57.1	17.4	96.5	89.5	97.7
6372	1942.2	57.0	147.6	58.5	17.8	98.2	91.0	99.4
6371	1941.9	56.4	146.0	58.1	17.7	99.9	92.4	
6370	1941.6	55.7	144.2	57.9	17.6	101.5	93.8	
6369	1941.3	54.9	142.1	57.7	17.6	103.3	95.4	
6368	1941.0	54.1	140.1	57.5	17.5	104.9	96.7	
6367	1940.7	53.6	138.8	57.1	17.4	107.0	98.5	
6366	1940.4	53.1	137.5	56.6	17.3	109.7	100.8	
6365	1940.1	52.6	136.2	56.1	17.1	110.9	101.8	
6364	1939.7	52.1	134.9	55.7	17.0	113.0	103.5	
6363	1939.4	51.7	133.9	55.1	16.8	114.9	105.2	
6362	1939.1	51.2	132.6	54.6	16.7	116.5	106.4	
6361	1938.8	50.7	131.3	54.1	16.5	119.2	108.7	
6360	1938.5	50.2	130.0	53.7	16.4	121.3	110.5	
6359	1938.2	49.7	128.7	53.1	16.2	123.7	112.5	
6358	1937.9	49.3	127.6	52.7	16.1	126.3	114.6	
6357	1937.6	48.9	126.6	52.2	15.9	128.5	116.4	
6356	1937.3	48.5	125.6	51.5	15.7	131.3	118.7	
6355	1937.0	48.1	124.5	50.8	15.5	133.6	120.6	
6354	1936.7	47.8	123.8	50.3	15.3	136.2	122.7	
6353	1936.4	47.4	122.7	49.9	15.2	139.0	125.0	
6352	1936.1	47.0	121.7	49.1	15.0	142.8	128.0	
6351	1935.8	46.5	120.4	48.5	14.8	144.8	129.6	
6350	1935.5	46.2	119.6	47.9	14.6	147.7	132.0	
6349	1935.2	45.8	118.6	47.4	14.3	151.0	134.6	
6348	1934.9	45.4	117.5	46.8	14.3	154.2	137.2	
6347	1934.6	45.0	116.5	46.2	14.1	157.5	139.8	
6346	1934.3	44.5	115.2	45.7	13.9	160.6	142.1	
6345	1934.0	44.1	114.2	45.3	13.8	164.6	145.3	
6344	1933.7	43.6	112.9	44.6	13.6	168.0	147.9	
6343	1933.3	43.1	111.6	44.3	13.5	172.1	151.1	
6342	1933.0	42.6	110.3	43.6	13.3	175.8	153.9	
6341	1932.7	42.1	109.0	43.3	13.2	180.1	157.3	
6340	1932.4	41.7	108.0	42.6	13.0	184.8	160.9	
6339	1932.1	41.2	106.7	42.0	12.8	188.8	163.9	
6338	1931.8	40.8	105.6	41.5	12.6	193.7	167.6	
6337	1931.5	40.3	104.3	41.0	12.5	198.2	170.9	
6336	1931.2	39.9	103.3	40.4	12.3	203.2	174.6	
6335	1930.9	39.4	102.0	39.9	12.2	208.0	178.1	
6334	1930.6	39.0	101.0	39.3	12.0	213.5	182.2	
6333	1930.3	38.6	99.9	38.7	11.8	219.1	186.2	
6332	1930.0	38.2	98.9	38.1	11.6	225.0	190.3	

Table 2. Variation of the predicted "Fish" production and plankton standing crop of Mono Lake as a function of the elevation of its surface.

Elevation (ft)	"Fish" production		Plankton standing crop	
	FP/U (kg/ha/y)	FP/L (1000 kg/lake/y)	PS/U (kg/ha)	PS/L (1000 kg/lake)
6428	2.04	47.4	33.1	769
6426	2.06	47.3	33.7	775
6424	2.07	47.2	34.2	779
6422	2.09	47.2	34.9	790
6420	2.11	47.3	35.6	799
6418	2.13	47.4	36.4	810
6416	2.15	47.6	37.2	825
6414	2.17	47.5	38.0	834
6412	2.19	47.9	39.0	852
6410	2.21	47.9	39.9	864
6408	2.23	48.0	40.8	877
6406	2.26	48.2	41.9	893
6404	2.28	48.2	42.9	906
6402	2.31	48.3	44.0	921
6400	2.33	48.2	45.0	932
6398	2.35	48.2	46.1	944
6396	2.38	48.3	47.4	962
6394	2.40	48.1	48.5	970
6392	2.43	48.1	49.8	987
6390	2.46	48.0	51.1	999
6388	2.48	47.8	52.3	1008
6386	2.51	47.8	53.9	1025
6384	2.52	46.6	54.2	1003
6382	2.55	46.4	55.7	1016
6380	2.56	45.6	56.6	1006
6378	2.56	43.9	56.4	967
6376	2.58	43.3	57.5	965
6374	2.55	40.3	55.8	883
6372	2.50	36.9	53.3	787
6370	2.52	36.3	54.1	781
6368	2.52	35.4	54.5	764
6366	2.55	35.0	55.7	766
6364	2.57	34.7	56.9	768
6362	2.60	34.4	58.4	774
6360	2.62	34.1	59.7	776
6358	2.65	33.8	61.3	782
6356	2.68	33.7	63.1	792
6354	2.72	33.7	65.2	807
6352	2.76	33.5	67.3	819
6350	2.79	33.4	69.6	832
6348	2.83	33.3	71.9	846
6346	2.87	33.1	74.3	855
6344	2.91	32.8	76.7	866
6342	2.95	32.5	79.1	872
6340	2.99	32.3	81.9	884
6338	3.03	32.0	84.8	896
6336	3.08	31.8	88.0	909
6334	3.13	31.6	91.5	923
6332	3.18	31.5	95.5	944

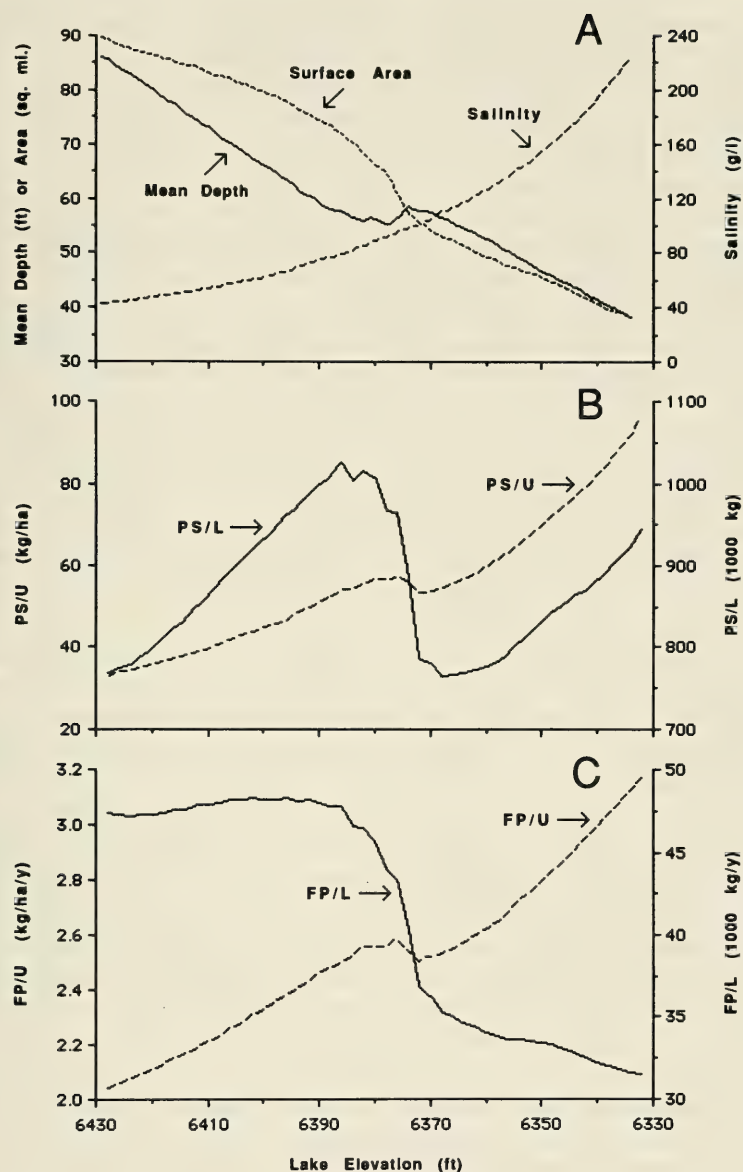


Fig. 2. Predicted variation of various properties of Mono Lake as a function of the elevation of its surface. PS/U = plankton standing crop per unit area; PS/L = plankton standing crop for entire lake; FP/U = fish production per unit area; FP/L = fish production for entire lake.

of change to be predicted for a change in Mono Lake's level. Nevertheless these results at least define a neglected issue in terms of a concrete example.

Analysis of the Predictions of the Models

The predicted variations of plankton standing crop and fish production are complex and quite different from each other. They illustrate well the potential influence of lake morphometry. They illustrate how the relative importance of lake area

(A) vs. lake mean depth (D) is a function of the power to which D is raised and thus can be very different for different biological variables.

Both FP/U and PS/U generally are predicted to increase as lake level drops (Figs. 2B, C). The general trend is reversed, however, when lake level drops from 6376 to 6372. That drop brings about declines of 3 percent for FP/U and of 7 percent for PS/U.

These 'bumps' in the curves are produced by a corresponding 'bump' in the curve relating mean depth to lake elevation (Fig. 2A). When lake elevation drops from 6376 ft to 6372 ft, lake maximum depth naturally decreases by 4 ft but lake mean depth *increases* by 3.2 ft (Table 1)! The apparent anomaly is due to the broad wave-cut platform, the Scholl terrace, that exists between roughly 6385 and 6370 ft around at least 75 percent of the lake's margin (Scholl et al. 1967; Pelagos 1987; Stine 1991). The lower boundary of this terrace is the so-called nick-point at 6368 ft. As the lake retreats across the gently sloped central portion of the terrace, lake volume decreases more slowly than does lake area and so mean depth increases.

Predicted values for total lake fish production (FP/L) and total lake plankton standing crop (PS/L) were obtained by multiplying FP/U and PS/U values by the corresponding values for lake area (A). The FP/L and PS/L curves are similar in some respects and dissimilar in others (Fig. 2)

Both FP/L and PS/L actually increase, PS/L quite substantially, as lake level drops from 6438 ft (its historic high stand, attained in 1919; Stine 1991) to about 6406 ft (FP/L) or 6386 ft (PS/L). For both variables the negative influence of declining lake area is more than offset by the positive influence of declining mean depth, over this range of lake elevations.

Then, starting at elevations around 6380 ft to 6384 ft, both FP/L and PS/L begin to decline rapidly with further declines in lake level. At 6380–6384 ft, lake level reaches the upper margin of the Scholl terrace, where the rate of decline in lake area per 1 ft drop in elevation begins to increase. The rates of decline of FP/L and PS/L are greatest as lake level drops from 6376 to 6372 ft and the negative influences of decreasing lake area and increasing mean depth reinforce each other.

The major differences between the FP/L and PS/L curves reflect the empirical fact that, at least among Rawson's (1952, 1955) Canadian lakes, FP/U is less strongly a function of mean depth than is PS/U. FP/U varies linearly with $D^{-0.7}$ and PS/U linearly with $D^{-1.5}$ (Eqs. 1, 2). Thus as lake level drops from 6428 to 6332 ft and mean depth decreases by 56 percent, FP/U increases only 56 percent while PS/U increases by 189 percent.

As lake level drops below 6370 ft, PS/L begins to increase again, the strongly positive effect of increasing shallowness overwhelming the negative effect of decreasing lake area (Fig. 2B). FP/L, on the other hand, continues to decline as elevation drops below 6370 ft, the influence of area dominating that of depth.

Despite the hypothetical nature of these calculations (Figs. 2B, C), I believe they provide insight into how the general productivity of Mono Lake, including the availability of invertebrate food for bird populations, may respond to changes in the depth and area of the lake. If depth or morphometry influences productivity at Mono Lake in a manner at all similar to that implied by Rawson's regression equations, the following conclusions are likely to hold for Mono Lake:

- 1) Over certain elevation ranges, a decrease in lake area will cause an increase in total lake productivity (TP/L);
- 2) TP/L may have increased for many years as water diversions caused lake level to drop from its 1940 pre-diversion elevation of 6416 ft;
- 3) With a particular change in lake level one index or component of TP/L may decrease (e.g., FP/L) while another may increase (e.g., PS/L);
- 4) TP/L will decline abruptly as lake level drops from 6380 to 6370 feet. A decline of 19 percent is expected solely from the 19 percent decrease in lake area; the 3.4 percent increase in mean depth and 19 percent increase in salinity both will increase the magnitude of this decline.
- 5) Salinity effects will depress the lefthand portions of the curves in Figs. 2B, C. Certainly no curve representing total invertebrate abundance or productivity will have a rising righthand limb.

These conclusions should be robust. They derive primarily from the particular morphometry of Mono Lake, and not from the specific form or coefficients of Rawson's equations. Many other functions representing a negative relationship between productivity per unit area and mean depth would yield the same conclusions.

Area and Brineflies

Though neither of the reports discusses the general importance of decrease in lake area, both did give specific attention to how the amount of hard, mostly tufaceous substrate in the littoral zone may decrease very abruptly when lake level drops from 6380 ft to 6370 ft. This hard substrate is strongly preferred by brine fly larvae and pupae over soft substrates. In the NAS report the expected reduction is repeatedly stated to be one of 40 percent (pp. 5, 189, 207), though the report's own Table 6.5 (p. 190) shows that the expected reduction would actually be either 61 percent or 57 percent, depending on whether the littoral zone is considered to extend to a depth of 10 ft or 30 ft. In any case, the magnitude of this decrease is more a consequence of the shallow water location of most of the lake's hard substrates than of the 19 percent decrease in lake area as it drops from 6380 ft to 6370 ft.

Need for a Reference Lake Level

A general and important question is, 'For any given past or future change in lake level, how do the salinity effects and lake size (area-depth) effects compare?' A great deal of attention is given in the NAS and CORI reports to defining at what salinity levels various negative impacts will "begin to occur." By the time such salinities are reached, however, marked reductions in the lakewide productivity or abundance of particular populations already may have occurred as a result of reduction in lake area. Naturally such assessments will depend on what lake level is taken as the baseline condition. This is another matter not explicitly addressed by the NAS and CORI reports, so a few comments on it may be appropriate.

The selection of a reference lake level must be partially arbitrary, but can also be partially objective. At least two lake elevations—6417 ft and 6376 ft—might

usefully be selected to define reference conditions. The former is the approximate mean lake elevation (June values) for the decade prior to the initiation of water diversions from the Mono Lake basin. We have, of course, no biological data for the lake during that decade. The second elevation, 6376 ft, is the approximate mean lake elevation for the decade (1977–1986, June values) prior to the establishment of the NAS and CORI panels. The charge to the NAS panel, and perhaps implicitly to the CORI panel, was to assess the consequences of lake level changes for “current wildlife populations” (NAS report, p. vii). Given the high year-to-year variability of Mono Lake populations and the unusual meromictic state of the lake at the time of the NAS and CORI panel deliberations, it seems reasonable to take as a reference point the mean lake level during the decade when large amounts of ecological data began to be collected at the lake. Arguments could also be advanced for using whatever lake level was just sufficient to keep coyotes off Negit Island. In any case it should be apparent that *the magnitude of effects of lowered lake cannot be estimated until a reference lake level is specified and a ‘baseline state’ of the lake ecosystem defined.*

Historical Perspective

Both the NAS and CORI reports provide inadequate analysis of the historical changes in the salinity and size of Mono Lake and their possible effects on the ecosystem. Very complete historical records of these variables were available to the committees. Their failure to give them more explicit consideration may explain why the possible effects of *future* changes in lake size were ignored and why the possible effects of *future* changes in salinity tended to be discussed in “plateau and threshold” terminology.

Both reports present graphs showing the full historical record for lake elevation, but neither report does this for salinity or lake size. For most populations at the lake, however, salinity and lake size will be the more important direct determinants of abundance. Lake elevation itself is of significance primarily as a determinant of size and salinity.

It was implicit in a NAS report (pp. vii–viii) that future changes, observed or projected, in Mono Lake populations were to be judged against the “current” status of those populations. However, the mandate to the NAS panel also called for consideration of “historic . . . populations levels . . . of all terrestrial and aquatic species” (NAS report, p. vii). Moreover, even if the sole objective were to estimate the future trajectory of the Mono Lake ecosystem, would not the best foundation for that exercise be an analysis of the historic trajectory of the ecosystem? In no way does the lack of validated models or the paucity of biological information for the system prior to the mid-1970’s reduce the need for explicit consideration of how past changes in salinity and lake are may have altered the Mono Lake ecosystem.

Birds and Their Food Supplies

A few further comments on this topic are warranted as it is a central one. From the viewpoint of conservation, an understanding of the factors influencing the productivities of algal and invertebrate populations is valuable primarily in allowing prediction of the food supply available to the birds using Mono Lake. We must then be able to understand how the bird populations would respond to

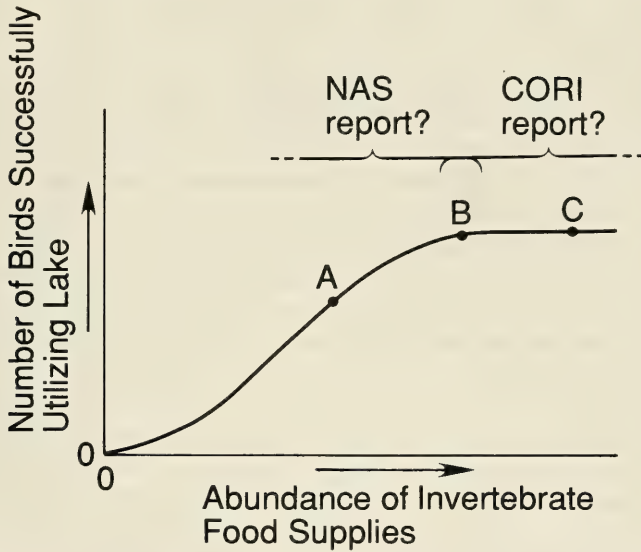


Fig. 3. A model of the relationship between invertebrate-eating birds and their food supply at Mono Lake. Points A, B, and C represent three possibilities for the present status of the system, as implied by the NAS and CORI reports.

changes in this food supply. And on that point somewhat uncertain but divergent conclusions seem to be reached by the two reports.

If brine shrimp and brine fly abundances were to decrease below a certain threshold, the birds that depend on them for food would be affected. Various effects can be imagined: reduced residence time at Mono Lake and movement to other lakes; reduced weight gain and survival on the lake; increased mortality following departure from Mono Lake, and so on. As J. Wiens points out in the CORI report (Appendix B), primarily in reference to the Eared grebe: "Unfortunately, we have no firm indication of the level of this threshold for Mono Lake bird populations with respect to either brine shrimp or brinefly abundances. It seems likely that, during recent years, summer and early fall shrimp and fly populations have been superabundant to the birds." The NAS report (p. 5), on the other hand, concludes that "If the lake fell to levels at which the birds' food sources were adversely affected, the bird populations would be reduced." In other words we are at or below the threshold now—*any* "adverse effect, *any* reduction in shrimp or fly abundance will have negative effects on the bird populations." "At" seems the intended idea. At least there is no intimation in the NAS report that more grebes and phalaropes, for example, would visit the lake if invertebrate densities were higher or the lake larger.

Both reports neglect, as indicated earlier, the role of lake size in determining total prey abundance and instead concern themselves only with the effects of salinity and, in the case of the brine flies, substrate availability. Discussion of bird energetics and physiology appropriately focuses on prey densities which in turn relate to per unit area productivity, but clearly for a given initial prey density a large lake can support birds in greater numbers and/or for a longer period of time than can a small lake. This is especially germane given the opinion of some (Cooper

et al. 1985) that grebe predation contributes to the autumnal decline of the brine shrimp population.

A threshold model for bird number-food supply relationships is depicted in Fig. 3, together with an indication of the differing viewpoints of the NAS and CORI reports. Note that a population at point B is consistent with the language of both reports. If the current status of a population at Mono Lake is accurately represented by point B or C, then the numbers that use Mono Lake are limited by some factor other than Mono Lake food supplies—such as habitat or nest site availability on the breeding range, for example. If the current status of the population is represented by point A, then food availability is indicated to be the prime factor limiting numbers of birds that use the lake. If there was a sustained increase in invertebrate productivity, we would expect eventually to see more birds using the lake.

Both reports failed to consider seriously that the current status of some Mono Lake bird populations might be represented by point A in Fig. 3. The omission was perhaps a consequence of the panels' failure to use explicit models or to wonder how bird food supplies might have been affected by the post-1940 25 percent decrease in lake area and 80 percent increase in salinity.

I am arguing neither that point A *does* represent the status of any particular species nor that the model in Fig. 3 *is* the most appropriate one, but only that those were and are important questions meriting deliberate consideration.

Conclusion

The criticisms offered here of the NAS and CORI reports have been put forward in an attempt to provide a broader perspective for the analysis of some aspects of the Mono Lake ecosystem. Discussions of the ecological, conservation, and water management issues will be clearer, more fruitful, and more appropriately focused if two things are done. First, all discussion of effects of changing lake levels should involve explicit models, even if these initially are only in the form of graphs with unscaled axes (e.g., Figs. 1 and 3). Such models are particularly needed 1) for expressing the presumed effects of salinity and lake size on the densities and sizes of invertebrate populations, and 2) for expressing the presumed effects of the invertebrate food supply on the number, behavior, and physiological state of certain bird species at Mono Lake. Second, explicit consideration should be given to how historical changes in salinity and lake area may have affected invertebrate and bird populations, even if, as is almost certain, firm conclusions cannot be reached on the topic.

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My understanding of the Mono Lake ecosystem has benefitted from discussions with many scientists who have worked at the lake, including especially T. Bradley, G. Dana, D. Herbst, J. Jehl, and J. Melack. Many of these discussions have taken place under the auspices of the Los Angeles Department of Water and Power and benefitted from the participation of LADWP's scientists, engineers, and lawyers. Calculation of the data in Table 1 was carried out by Randall Neudeck of LADWP. N. Sjaarda and J. Ferrara prepared the figures. T. Hammer, D. Herbst, J. Jehl, P. Richerson, and two anonymous reviewers are thanked for their comments on the manuscript.

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Mass Movement and Seacliff Retreat along the Southern California Coast

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Abstract.—Seacliff retreat is a complex response between the magnitude and frequency of interactive processes, primarily mass movement and marine erosion, and the properties of coastal terrain. Mass movement occurs on or above seacliffs when resisting forces are overcome by driving forces, especially when the shear strength of cliff-forming materials is reduced by absorption of water or when shear stress is increased by loading, vibration, or removal of toe support. Many landslides of the Malibu and Palos Verdes coasts are prehistoric features reactivated in recent time by heavy winter rains or by human impacts on slope hydrology and buttressing. Marine erosion removes both solid rock and mass-movement debris from the cliff base by hydraulic forces and abrasion, the efficiency of which reflects hydrodynamics and resistance properties. Seacliff retreat is episodic and site-specific. Harder rocks, like Cretaceous sandstone, tend to erode more slowly than softer Quaternary deposits. Recent retreat rates along the San Diego and Santa Barbara coasts vary from negligible to 0.5 m/yr. The nature and rate of seacliff retreat were largely ignored during the region's early development but have recently been investigated quantitatively and incorporated into coastal management plans.

Seacliffs characterize some 60% of southern California's 450 km mainland coast between Point Conception and the Mexican border, and also largely encircle the offshore islands. Some seacliffs are actively retreating from the combined impact of marine erosion and mass movement, notably between Point Conception and Santa Barbara and between San Onofre and San Diego. Elsewhere, seacliffs are now often protected from marine erosion by coastal highways, for example between Santa Monica and Malibu, but mass movement still occurs.

Scientific interest in coastal erosion has a long history but in southern California threats to human life and livelihood posed by mass movement and seacliff erosion have recently expanded this interest into issues of practical concern. Southern California's early growth focused mainly on low-lying coasts with space for harbors and access to the interior. Subsequently, however, urban growth spread onto and below coastal cliffs, often with little regard to the physical hazards of such development. Government agencies and private citizens alike combined in ignorance to develop inherently unstable coastal slopes and to trigger instability by changing the physical properties of previously stable areas. Early attempts at coastal-zone management were largely ineffectual and it was not until 1976 that passage of the California Coastal Act provided the formal framework for a more rational management of the coast. Nevertheless, a legacy of earlier manipulation of seacliffs and coastal slopes remains.

Coastal-zone management provides the societal context within which most scientific and engineering investigations of seacliff erosion and coastal landslides are now conducted. The research emphasis of the past three decades has moved away from purely descriptive studies towards an understanding of the processes involved in seacliff retreat and the acquisition of quantitative data regarding rates of change. These themes are the focus of this paper.

Processes of Seacliff Retreat

Seacliff retreat is episodic, site specific, and related to the magnitude and frequency of the processes involved and the resistance of the cliff materials. Two groups of processes are primarily involved, namely mass movement and marine erosion. Aided by weathering and flowing water, these processes are commonly interdependent: marine erosion undermines the cliff base and thereby triggers mass movement from above; conversely landslides gravitate to the shore where their toes are eroded by wave action, in turn generating further instability. The cliff profile is the product of an equation integrating these processes, tending to steepen towards the cliff base where marine erosion is dominant but declining to gentler, if unstable, slopes where mass movement and subaerial erosion prevail.

Ancillary factors also affect seacliff retreat. For example, water levels that rise or fall beyond the normal range of tide and wave activity may influence groundwater tables, and thereby affect weathering and mass movement. This is a troublesome issue around the Great Lakes and may cause significant long-term changes along open coasts exposed to regional and global sea-level changes.

Earthquakes and tsunamis may also trigger seacliff collapse. The Loma Prieta earthquake (M 7.1) of 17 October 1989, which ruptured a 40 km segment of the San Andreas fault zone and caused much destruction in the San Francisco Bay area, generated extensive cracking and collapse of coastal bluffs in the Santa Cruz area (Plant and Griggs 1990). Similar cliff collapse occurred in the region during the 1865 earthquake (?M 6.5) and the M 8.3 San Francisco earthquake of 1906. Along the southern California coast, evidence for recent earthquake-induced cliff collapse is less dramatic. The M 6.3 Long Beach earthquake of 1933 caused cliff collapse at Malaga Cove and elsewhere around the Palos Verdes Hills. In March 1968, a moderate earthquake near San Diego caused part of Sunset Cliffs to slump shorewards. In February 1971, the M 6.5 San Fernando earthquake triggered numerous rockfalls along the Malibu coast, although the epicenter lay over 40 km north. It is likely that some prehistoric coastal landslides were activated by earthquakes for which we have no precise record. Furthermore, earthquake-induced fracturing may promote mass movement weeks or months later. Certainly, the probability of earthquake-induced cliff collapse and mass movement remains high along the southern California coast, especially along the active Malibu Coast, Newport-Inglewood and Rose Canyon fault zones where Holocene deposits record relatively recent dislocations (Fig. 1).

Mass Movement

Mass movement, the downslope motion of earth materials under the influence of gravity, is a global phenomenon but is often prominent where coastal slopes are undermined by erosion or modified by other processes, including human activity. Movement is seldom attributable to a single identifiable cause; in most

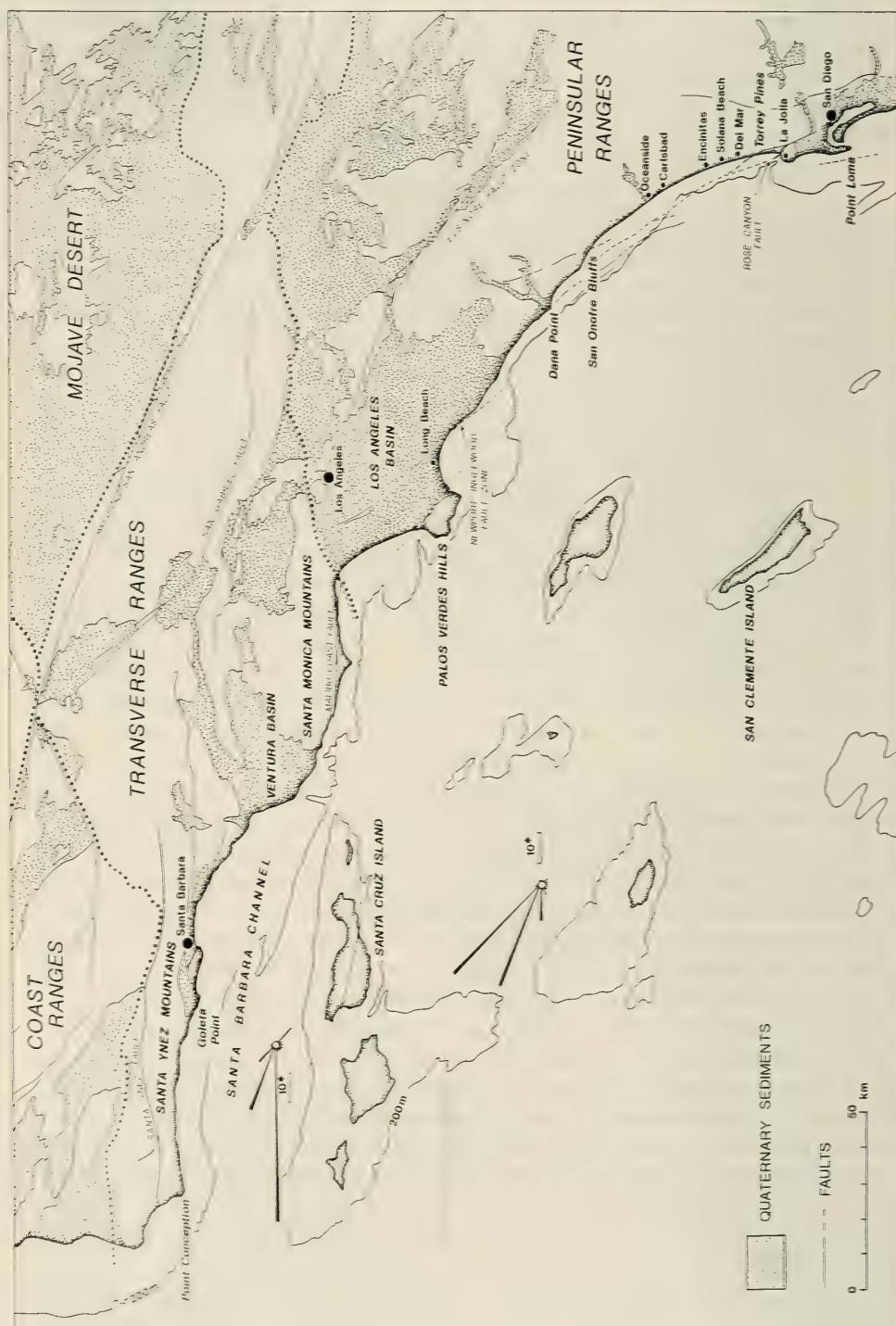


Fig. 1. The southern California coast showing locations discussed in text. *Roses for deep-water wave power ($\times 10^{-11}$ joules/yr/m) are shown north and south of Santa Cruz Island.



Fig. 2. Landslide terrain in Miocene shales, northern Santa Cruz Island. As the hummocky landslide complex rotates upward at the shore, it is cliffed by marine erosion (A. R. Orme, May 1970).

cases, several causes exist simultaneously and the last is merely a trigger that sets in motion a mass that was already close to failure (Fig. 2). The following discussion outlines the principles of mass movement, with examples from the Malibu and Palos Verdes coasts of Los Angeles County.

Simply stated, mass movement occurs when the forces tending to cause movement overcome the forces resisting change. In geotechnical terms, this relationship is defined by the Safety Factor, namely the ratio of resisting forces to driving forces along a potential failure surface (or the ratio of resisting moments to driving moments about a point). Ideally then, failure is represented by a Safety Factor less than one, and stability by a value greater than one. Resisting forces are in turn defined in terms of the available shear strength of a material, whereas driving forces are expressed as average shear stress.

A body at rest on a slope resists movement because its shear strength exceeds the shear stress to which it is exposed. Intact shear strength (S) is a function of three variables: the physical and chemical cohesion (C) that bonds material; the body's weight directed as a force normal to the slope, or effective normal stress (ϕ); and the angle of internal friction (ϕ) which determines the threshold to which a body will resist movement against another body. Thus intact shear strength is defined as $S = C + \sigma \tan \phi$. Low shear strength is a fundamental property of some earth materials, attributable to their composition, texture or structure. Thus platy clay minerals, cohesionless sands, fractured bedrock, seaward-dipping strata, and alternating permeable and impermeable beds all favor low shear strength.

Shear strength may be further reduced by weathering and the addition of water. Weathering of coastal cliffs includes the physical disintegration of granular rocks

by salt-crystal growth accompanying the evaporation of spray, the hydration of clay minerals such as montmorillonite and illite which swell and lose cohesion in the presence of water, the solution of intergranular cements, and other processes that weaken physicochemical bonds between minerals.

Because most weathering reactions involve water in one form or another, the relationship between water and shear strength may also be viewed in a broader context. Adsorption of water reduces the physical cohesion of earth materials, which in moist but unsaturated materials is a suction force drawing particles together by capillary tension. The addition of water increases porewater pressure (u) which reduces normal stress between particles by forcing them apart. Restating the above equation as $S = C + (\sigma - u) \tan \phi$ summarizes this reduction in shear strength. In essence water creates a buoyant force in saturated materials that reduces effective intergranular pressure and friction. Such a reduction in the effective shear strength of either discrete seams or large masses of earth eventually promotes mass movement.

Apart from reduced shear strength, mass movement is also triggered by increased shear stress attributable to the removal of support or to loading, vibration, or tilting. Removal of support from the toe of a slope is readily accomplished by marine erosion or road construction. Furthermore, once mass movement occurs, instability tends to be self-perpetuating until some balance is restored. This is because the mass that has shifted is no longer able to support the slopes above. Loading or surcharge by absorption of water into upslope materials creates an imbalance in the static stress field, although increased lateral stresses are partly compensated by increased normal stresses. Loading may also occur from such human activities as construction of cliff-top buildings and emplacement of fill. Vibrations caused by earthquakes, traffic or pounding waves also increase shear stress. Prolonged ground shaking during earthquakes may decrease shear strength to the threshold of liquefaction whereby water-saturated materials with little or no cohesion are transformed into fluid masses. A progressive increase in slope angle through tectonism also increases shear stress over the longer term, as shown along the south flanks of the Santa Monica Mountains.

The role of vegetation in slope stability is much debated. Removal of native plants may reduce near-surface shear strength by removing mechanical root strength and water transpiration by plants. This may increase the likelihood of shallow slope failures but has minor impact where failure surfaces are relatively deep. In the Portuguese Bend and Abalone Cove landslides, however, roots of eucalyptus and pepper trees penetrate to depths exceeding 30 m and extract significant amounts of groundwater. Plants also represent biomass whose forces are directed both normal to and parallel with the slope. Vegetal loading and unloading may thus yield complex responses.

Earth materials adjust initially to increased shear stress by mass straining without rupture, but continued stress may eventually exceed the material's shear strength, the safety factor is reduced to one or less, and mass movement occurs.

Mass movement is thus favored by reduced shear strength or increased shear stress, but the response is usually complex and not easily summarized. Nevertheless, mass movements are commonly categorized as falls, flows, and slides. Falls of earth and rock are common where seacliffs are undermined by marine erosion or where shear strength is compromised by groundwater fluctuations or

vibration. The resulting accumulation of talus at the cliff base may protect the cliff until removed by waves, which may explain why high cliffs that produce abundant talus often retreat relatively slowly.

Flows are characterized by non-Newtonian plastic behavior of sediment-water mixes involving relatively high sediment concentrations. Although often generalized as debris flows, they may involve a range of materials from mud to rock supported by matrix strength or dispersive pressure, or in more fluid flows by escaping pore fluids. Flows vary in velocity from mm/day to m/s, with swift flows typically coinciding with heavy rains which rapidly mobilize unconsolidated debris.

Notwithstanding the importance of falls and flows, landslides are probably the most impressive forms of mass movement found along the southern California coast, mainly because they often involve large, deep-seated, and seemingly inexorable movements of mass towards the shore. Landslides are favored by certain structural and stratigraphic conditions, notably by seaward-dipping strata comprising permeable and impermeable beds and by fractured and pulverized bedrock. They are underlain by one or more slide planes, which may be translational as in block glides, or rotational where failure occurs along arcuate surfaces. Large landslides are characterized by distinctive surface features, notably steep headscarps, secondary scarplets, pressure ridges, closed depressions, hummocky ground, and lobate toes, and by such stratigraphic or structural features as rotated and offset bedding, disrupted strata, clastic injections, and various slide planes.

The relationship between mass movement and water is complex. Shallow slides and flows often occur during or shortly after heavy rains. Deep-seated landslides, however, may occur days or even months after major rains because it takes that long for water to infiltrate fractures and pores, and modify shear strength at depth. Such effects are less delayed after prolonged dry periods because cracked earth readily accepts rain and runoff. Human activity such as lawn watering, septic-tank seepage, and road drainage may promote movement at any time by raising groundwater levels and modifying shear strength. Further, because so much landslide terrain has been inherited from wetter Pleistocene climates, such human activities may recreate the groundwater hydrology and stress fields of the Pleistocene, pushing terrain of marginal stability over the threshold into renewed landslide activity.

Bluffs along the Malibu coast are prone to mass movement for several reasons. They are formed either in poorly consolidated alluvial and colluvial deposits or in pervasively fractured and weathered bedrock within the complex Malibu Coast Fault (Fig. 3). These slopes failed episodically during Pleistocene times and were further destabilized by seacliff erosion following the Flandrian marine transgression. More recently, cliff-top developments involving vegetation change, grading, road drainage, septic tanks, and landscape watering have greatly changed hillslope hydrology. Such development began at Santa Monica in 1875, at Pacific Palisades in 1921, and at Huntington Palisades in 1924. Meanwhile, early railroad and later highway construction at the cliff base has greatly reduced buttressing slope support. For example, the Hueneme, Malibu, and Port Los Angeles Railroad was opened in 1908 over and below 24 km of coastal bluffs from Malibu to near Little Sycamore Canyon. Much of this route was later absorbed into the Roosevelt Highway (now Pacific Coast Highway) which, after more than six years of con-



Fig. 3. Landslides and seaciffs of the Malibu coast, southern California. Landslides based mostly on McGill (1973) and Yerkes and Campbell (1980).



Fig. 4. Via de las Olas Landslide, Pacific Palisades. On 31 March 1958, this abandoned but persistently unstable seacliff failed after heavy rains. Rather than remove the toe of the landslide complex, the Pacific Coast Highway was relocated seawards (Spence Collection, UCLA, April 1958).

struction, was opened in June 1929 for through traffic between Malibu and Oxnard. Farther east, a public coastal highway had been declared as early as 1906. The widening and westward extension of this highway in the 1920s required some two million m^3 of rock debris to be removed from the cliff base between Santa Monica Canyon and Point Mugu for use as road bed and seaward protection. Added protection was afforded by 24 groins, 30,000 m^3 of riprap, and several long seawalls. Understandably, the coastal bluffs were thrown into disequilibrium from which they are still recovering. A few examples will illustrate this point.

Pleistocene fanglomerates forming the 30–40 m Santa Monica and Huntington Palisades have failed frequently in recent decades, casting cliff-top property onto the coast highway. Railroad construction between Colorado Avenue and a wharf off Potrero Canyon in 1891 initiated disruption of the cliff base and was followed by several phases of highway construction. After one widening phase, cliff failures in 1932 led to the building of an extensive surface and sub-surface drainage system that included an attempt to dry out the cliffs using heaters placed in tunnels!

Neogene sediments within the Santa Monica fault zone have a history of recurrent failure. In March 1958, a perpetually troublesome area on Via de las Olas failed again after heavy rains, carrying 600,000 m^3 of debris seawards (Fig. 4). The coast highway was subsequently relocated around this slide area and protected

by groins. Northwest of Temescal Canyon, a large prehistoric landslide remains unstable, occupied appropriately by mobile homes.

Farther west, Pacific Coast Highway is destined to be a continuing focus of mass movement. One bedrock bluff east of Piedra Gorda Canyon, the Big Rock Landslide, developed arcuate tension cracks above the highway in winter 1979, causing dangerous rockfalls and threatening to shift two million m³ of rock seawards. The rockfall zone rose 80 m at 45–60°, above which a zone of tension cracks extended a further 120 m across 15–35° slopes. These cracks were opening at a rate of 3 cm/day in April 1979. The highway authority subsequently removed about 115,000 m³ of material, terraced the cliff face, and installed a protective barrier at the base, but complete stabilization needs greater slope reduction.

Big Rock Mesa Landslide, between Piedra Gorda and Las Flores canyons, is a 100 ha prehistoric complex that went largely unrecognized until renewed movement began in the 1970s, by which time nearly 300 homes occupied the slide area (Fig. 5). Although some homes were built as early as 1928, most development occurred during the 1950s and 1960s in conjunction with private sewage-disposal systems. Although high initial groundwater levels had been lowered by pumping from wells, such pumping was later abandoned. Septic-tank effluent, landscape watering, and heavy winter rains in 1978 and 1980 subsequently recharged groundwater aquifers, reduced shear strength, and triggered movements which peaked at 3–5 mm/day in the early 1980s. The area is geologically complex, involving several rock formations of varying properties lying between the Las Flores and Malibu Coast faults (Yerkes and Campbell 1980). Rotation of this complex along a series of slide planes, the lowest descending below sea level, opened a 3–6 m high headscarp and several subsidiary scarps, and caused significant deformation within the slide and along Pacific Coast Highway. Stabilization efforts, begun in 1983, involve using vertical wells between 25 and 170 m deep and hydrauger drains to dewater the landslide mass (Michael 1986). These methods have achieved some success in lowering the piezometric surface during recent relatively dry years, but whether dewatering will be as effective under wetter conditions is uncertain. Furthermore, a larger area than that involved in recent movement appears capable of shifting seawards south of the Las Flores Fault.

Large landslides also occur along the south flanks of the Palos Verdes Hills (Fig. 6). Some are recent but most are ancient features reactivated in historic times. As these landslides reach the shore, their toes are eroded by wave action, which in turn triggers further movement. The area is underlain by Neogene marine sediments and volcanics of the Monterey Formation which, because they have been arched upwards along the NW-SE anticlinal axis of the hills, dip seawards to the south at between 6° and 25°. Of these rocks, the Altamira Shale (15.5 to 13.5 Ma) is important from the landslide perspective because it contains beds of volcanic tuff from 1 cm to more than 20 m thick. These tuffs decompose to bentonite whose high montmorillonite clay fraction and fine texture produce a soft plastic material which swells up to eight times its original volume in the presence of water. These 'swelling clays' thus suffer reduced shear strength and also provide effective impermeable barriers to groundwater movement, thereby increasing hydrostatic pressures in overlying rocks.

The Point Fermin Landslide initially separated from its arcuate headscarp in 1929 and was reactivated in 1940. It first moved seawards as a block glide over



Fig. 5. Big Rock Mesa Landslide, Malibu. This 1933 photograph predates most development in the area and clearly shows hummocky landslide terrain forming a broad bench above Roosevelt (now Pacific Coast) Highway. The crest of the primary headscarp rises along the right side of the photograph, above the track (Spence Collection, UCLA, February 1933).

a south-dipping bentonite seam 1–2 cm thick. Its eastern portion later moved southwest and broke into several rotational blocks. Because its toe is undermined by wave vibration and erosion of 30 m high seacliffs, this portion continues to move episodically seaward along a 300 m front (Fig. 7).

The South Shores Landslide extends as a glide block some 1200 m from an elevation of 275 m to just below sea level. It reaches the coast along a 900 m front in cliffs up to 60 m high (Fig. 6). Based on a ^{14}C age of $16,200 \pm 240$ years BP for alluvium trapped by the landslide, initial movement evidently followed the emergence of the latest Pleistocene marine terrace (Ray 1982). Although presently inactive, the area has the potential for reactivation. Part of a housing tract lies within the landslide, having been under construction in 1956 when government agencies first alerted the developer to the hazard.

Farther west, the 400 ha Portuguese Bend landslide complex occupies a broad amphitheater extending for 2 km from near the crest of the Palos Verdes Hills to the sea (Fig. 6). It is a deep-seated prehistoric complex that probably developed during and after the middle to late Pleistocene emergence of the marine terraces by which it is flanked (Woodring et al. 1946). However, it is its partial reactivation in recent decades that has aroused most interest (Merriam 1960; Vonder Linden 1972; Ehlig 1986). Active components range from massive glide blocks with typical headscarps to smaller rotational slumps and hummocky ground with numerous scarplets and small grabens. Furthermore, an arcuate graben that extends

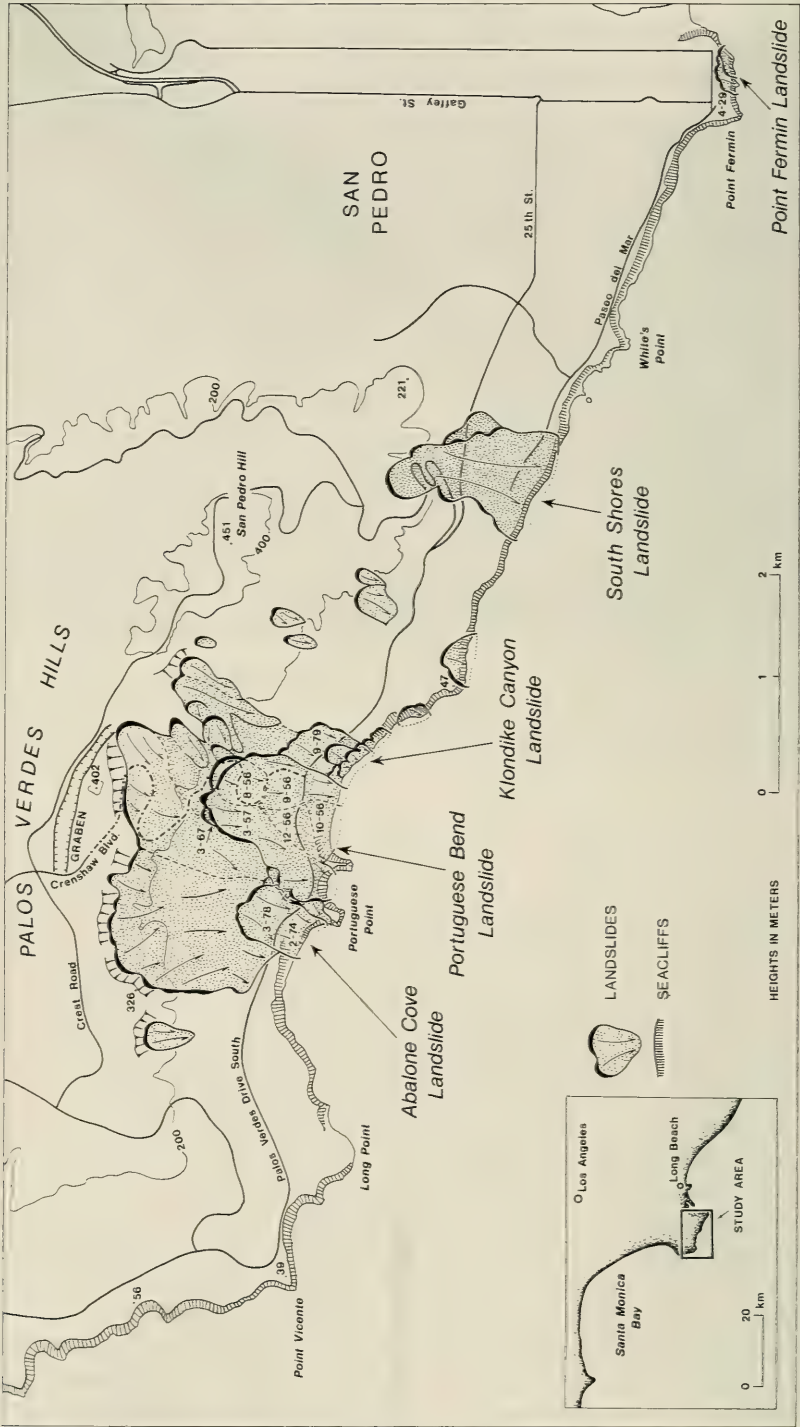


Fig. 6. Landslides and seaciffs along the south coast of the Palos Verdes Hills. Month and year of reactivation are shown for some landslides (e.g., 8/56). Based on references in text and author's observations.



Fig. 7. The Point Fermin Landslide in February 1941, showing the relationship between mass movement and marine erosion. Reactivation in 1940-41 occurred within the framework of the larger 1929 landslide whose subdued headscarp is visible between buildings towards the top right (Spence Collection, UCLA, February 1941).

for more than 1500 m along the crest of the hills may also have formed from landslide activity, perhaps aided by tectonic extension (Merriam 1960; Ehler 1986).

Three areas of reactivation are commonly recognized, namely the Klondike Canyon, Portuguese Bend, and Abalone Cove landslides. Renewed movement of the Klondike Canyon Landslide was recognized in 1979. The mass is underlain by several seaward-dipping tuff beds, including the 15 m thick Portuguese Tuff, and reaches the coast in eroding seacliffs up to 50 m high (Kerwin 1982).

The 110 ha Portuguese Bend Landslide began renewed movement in August 1956, again in association with seaward-dipping tuffs and low-angle slide planes, some of the latter overlying inactive landslide material. Los Angeles County was later found legally responsible for this reactivation, having loaded the headscarp area with roadfill during the extension of Crenshaw Boulevard (Ehlig 1982). Increased groundwater levels associated with home construction and water-line breakage may have aggravated the problem. Certainly, water has since come to be regarded as the primary trigger of continuing activity. Initial movement was a dramatic 7–10 cm/day and, although this later slowed, cumulative lateral displacement of 80–240+ m has occurred since 1956. Towards the coast, the slide mass thrusts upwards at a steep angle against the 30 m thick basalt sill that buttresses Portuguese Point and Inspiration Point. In the broad cove to the east, reinforced concrete caissons were emplaced in 1957 to stop movement but these

have since been displaced up to 140 m! Some 134 homes within the slide area were destroyed or removed as a consequence of these movements.

The 32 ha Abalone Cove Landslide began moving along a 360 m front west of Portuguese Point in 1974 and, after heavy winter rains in 1978, movement accelerated and also extended northwards across Palos Verdes Drive South. Between 1977 and 1981, this mass shifted 8 m laterally and dropped 1 m. Subsequent research focused on the driving forces generated by mass downdrop along listric normal faults within the mostly translational slide, on buoyancy imparted by rising groundwater caused by infiltrating rainfall and on-site sewage disposal, and on distinguishing between the forces required to initiate motion and the much lesser forces needed to sustain movement (Ehlig 1982).

Over the past decade, dewatering systems involving vertical wells and improved surface drainage have retarded and locally stopped movement of the Portuguese Bend and Abalone Cove landslides. Transfers of earth material have also served to reduce driving forces. After 1980, however, winters in the area have experienced either less than normal rainfall or more evenly spaced storm events, and it is likely that significantly heavier rains might again accelerate movement or that increased storminess and seacliff erosion at the coast might remove buttressing mass.

The landslides of the Malibu and Palos Verdes coasts illustrate well the relation between mass movement and seacliffs past and present. Elsewhere, this relationship is more directly expressed at the cliff face, notably along the Santa Barbara and San Diego coasts.

Marine Erosion

Marine erosion of seacliffs involves the removal of materials from the cliff base by hydraulic forces and abrasion. Of the hydraulic forces, compression occurs when air in cracks and fissures is confined by waves hitting the cliff face, whereas tension occurs as water recedes and the compressed air expands rapidly. Cavitation, the growth or collapse of air cavities in a liquid caused by isothermic pressure changes, occurs when water flows at high velocity against a cliff face. These interacting forces can, if repeated sufficiently often, promote rock fatigue and thereby cause separation and collapse of cliff materials.

Abrasion results from both the rasping effect of debris-laden waters near the cliff base and the artillery effect of cobbles and boulders shot at the cliff face by waves. Abrasive materials are provided by prior seacliff retreat, by nearby rivers, and by erosion of offshore reefs and shelf deposits. If this supply is large, cliff erosion will be inhibited by beach accumulation, but if supply is countered by debris removal offshore then cliffs will retreat across a widening abrasion platform that slopes seawards at 1° to 6°. In hard rocks, a wave-cut notch may develop near the cliff base, a more-or-less horizontal sawcut in the zone of most effective abrasion.

The efficiency of hydraulic forces and abrasion reflects hydrodynamic influences, coastal morphology, and rock resistance. Wave climate is especially important and is in turn affected by meteorological conditions and coastal orientation. North of Point Conception, 30–50% of deep-water swells approach the California coast from the north west, generated by prevailing northwesterly winds related in summer to strong clockwise outflows of air from the Hawaiian high pressure cell and

in winter to counterclockwise flows towards Aleutian low pressure cells or associated with the eastward progression of cold fronts across the state. South of Point Conception, however, changing coastal orientation and offshore islands cause 70% of swells to pass up the Santa Barbara Channel from due west and 80% of swells to approach Los Angeles from west-southwest (Fig. 1). Here also, more southerly swells are set up by late summer hurricanes off western Mexico, by storms in the Southern Hemisphere winter, and by local winter depressions passing along more southerly tracks. Thus the San Diego County coast is moderately exposed to northwesterly swells but no part of the coast is protected from more southerly swells and storm seas. Erosion potential is also affected by periodic sea-level changes related to tides and lower frequency events. For example, high sea levels are a common manifestation of El Niño-Southern Oscillation events which recur every four to seven years owing to the relaxation of westward-blowing trade winds in the tropical Pacific. Thus storm effects during the El Niño winter of 1982–83 were superimposed on sea level 10 to 20 cm above the long-term mean (Flick and Badan-Dangon 1989).

The depth and morphology of the nearshore bottom also influence waves approaching the cliff base. In deep water, the wave phase velocity (C_o) is primarily related to wave length (L) and water depth (d) is unimportant. Thus $C_o \approx \sqrt{gL}$, where g is the acceleration of gravity. As water depth decreases to about $0.25 L$, however, deep-water waves begin to feel bottom and deform. When water depth lessens to $0.05 L$, shallow-water waves occur whose phase velocity (C_s) is primarily controlled by depth. Thus $C_s \approx \sqrt{gd}$. As water depth and phase velocity decrease, wave length decreases but, in order to transfer energy at a constant rate, wave height (H) increases. Now, because water depth varies between wave crest and wave trough, shallow-water wave velocity is greater at the crest and this disparity increases shorewards as wave steepness (H/L) increases. Thus the wave crest advances towards the preceding trough, wave form becomes more asymmetric and unstable, until the wave breaks when $H \approx 0.8d$. On the mainly steeply shelving coast of California, ocean swells approach farther inshore and transform to shallow-water waves over shorter distances than on gently shelving coasts, although the latter occur off major river mouths.

Breaker types, the form in which waves project energy shorewards, are also related to wave height and water depth. Surging breakers are favored by steep nearshore slopes, break close inshore, and generate short but strong onshore currents (swash) that can throw much water and sediment against a cliff face. In contrast, spilling breakers are favored by gentle nearshore slopes, break far from shore, and generate relatively weak swash and backwash. On rocky coasts with wide abrasion platforms, spilling breakers rarely accomplish much cliff erosion. Plunging and collapsing breakers are intermediate forms whose erosional impact on rocky coasts depends on platform width and tidal stage. Negative feedback characterizes these relationships: as cliffs retreat so the platform widens and waves expend more energy on the latter and less on the former. Thus cliff retreat rates tend to slow with time; then, as positive feedback, unremoved talus further protects the cliff base from erosion.

The efficiency of marine erosion also depends on the composition, consolidation, and structure of cliff-forming materials through the influence of these properties on chemical strength and mechanical resistance. Except for limestones,

which do not outcrop much along the California coast, rocks of simple composition dominated by quartz and clay minerals are less prone to chemical weathering than complex igneous and metamorphic rocks whose feldspars and micas are especially liable to decay. All coastal rocks are, however, prone to mechanical weathering, especially wetting and drying which causes clay minerals to expand and contract, and granular disintegration caused by salt crystallization from spray. In terms of consolidation, rocks produced by fusion are strongest, those consolidated by cementation of middling strength depending on the cement (Ca, Fe, Mn, etc.), while compaction produces relatively weak rocks. Thus the igneous and metamorphic rocks of the offshore islands form bolder coasts than the Cenozoic sedimentary formations lining much of the mainland shore. Among the latter, sandstones tend to be more resistant than siltstones, siltstones stronger than shales (Fig. 8).

Rock strength is further compromised by structural weaknesses imposed by bedding planes, joints, faults, cleavage, and foliation, the continuity and spacing of which influence cliff form. Thus an otherwise mechanically strong rock like basalt may be quarried along joint planes formed when the lava cooled. Some coastlines are strongly fault-controlled, notably the northeast slopes of San Clemente Island and the Malibu coast. Structural attitudes of sedimentary rocks also influence cliff form: where rocks dip landward, or seaward at 10° or less, seacliffs are often vertical, as seen along the San Diego coast; where seaward dips exceed 30° , for example west of Santa Barbara, seacliffs may coincide with bedding planes.

Bioerosion also aids cliff retreat. Rocks along the California coast that are heavily infested with the rock-boring clam *Penitella penita* are eroding at about 1.2 cm/yr, whereas rocks not thus afflicted are retreating at only 0.05 cm/yr (Evans 1968). The date mussel, *Lithophaga plumula*, which bores by secreting acid, may be seen attacking rocks in sheltered bays. Bioerosion is also associated with other pelecypods as well as grazing limpets, chitons, echinoids, sponges, and endolithic algae. Coalescence of shallow intertidal pans along the San Diego coast has been attributed to solution of carbonate cement during the night-time development of low pH by respiration of plants and animals, especially the snail *Littorina planaxis* (Emery and Kuhn 1980).

Rates of Seacliff Retreat

The rate of seacliff retreat is a function of the relationship between marine erosion and cliff resistance over time. Mass movement complicates this relationship, either by achieving in a few minutes what it may take marine erosion many years to accomplish, or alternatively by protecting the cliff base with debris. Some early investigators concluded that seacliff erosion in southern California was negligible (Shepard and Grant 1947). Recent research reveals a more variable pattern with some cliffs retreating rapidly while others experience little measurable erosion.

The Nature of the Evidence

Evidence for seacliff erosion rates occurs in two main forms: documentary materials and field data. Documentary materials include old maps, marine charts, photographs, tax assessments, highway surveys, railroad records, and newspaper accounts. From these it may be possible to compare cliff location from one time to another, and thereby derive an erosion rate. For example, the United States



Fig. 8. Seacliffs at Torrey Pines in 1928. The lower, continuous, near-vertical 30 m seacliff is reinforced with Eocene sandstone; the dissected middle cliff developed in Eocene siltstone and shale mantled by Pleistocene colluvium; and the crest capped at around 100 m above sea level by other Pleistocene sediments. The cliff base is clear of talus, indicating effective marine erosion at this time (U.S. Navy, June 1928).

Coast Survey (now the National Ocean Service (NOS) of the National Oceanic and Atmospheric Administration (NOAA)) was established in 1807 and began charting the California coast in 1851. Likewise, the United States Geological Survey (USGS), founded in 1879, began producing maps based on field survey in the 1890s, since updated by further surveys and photogrammetry. Earlier surveys from the Spanish and Mexican periods, and sea charts compiled by English, Spanish and Russian navigators, though interesting, provide little basis for precise measurement. On the other hand, railroad right-of-way surveys date from 1853 and sometimes contain notations concerning when and where tracks were damaged by erosion and landslides. Map data should always be treated cautiously because of potential errors associated with original survey and subsequent cartography, especially problems of scale and generalization.

Oblique ground photographs of the coast are available from the mid-nineteenth century and may offer useful comparative data, although measurement is often difficult. Aerial photographs became generally available in the 1920s and, with sufficient resolution, offer valuable records of coastal change. The problem of using old maps and aerial photographs was highlighted by a study of shoreline locations conducted by NOS and the United States Army Corps of Engineers (1985) which compared maps from 1852 onwards and 1982 aerial photographs of the coast between Palos Verdes and the Mexican border. Problems of defining

'mean high water line' and cliff base resulted in maps at the 1:24,000 scale with an accuracy of only ± 12 m. Not surprisingly, long stretches of cliffed coastline revealed little measurable change, although northern San Diego County showed significant cliff retreat while bluffs at Portuguese Bend seemingly shifted 150 m seaward between 1870 and 1982. Documentary methods for evaluating historical coastal erosion have been demonstrated for the Encinitas area by Fulton (1981).

Field data are acquired by repeat surveys that establish the distance between some fixed point and a seacliff. Triangulation stations established by survey agencies are valuable in this context. Where existing survey markers are lacking, erosion pins may be emplaced. Cultural features such as roads, buildings and even dated inscriptions may offer similar baseline data. Emery (1941) observed that 163 dated inscriptions averaging 3 mm deep on sandstone cliffs at La Jolla were erased in 6 years, for a mean retreat rate of 0.0005 cm/yr. He therefore concluded that other processes, such as rock falls following wave quarrying, must explain higher rates of cliff retreat. Microerosion meters designed to measure shore-platform denudation have proved less useful on vertical cliff faces. Whatever survey point is used, it is important to establish what is being measured because, with marine erosion and mass movement both involved, the cliff base may retreat at a different rate from the cliff top.

Spatial Variability

Seacliff retreat varies considerably in space and time. Spatial variability is well illustrated between Dana Point and Point Loma, a 100 km stretch of coast wherein exposure to wave attack is similar. Dana Point is formed in resistant Miocene breccia, but to the north less competent shales of the Monterey Formation favor landslides while for 14 km southward poorly cemented Miocene siltstone and sandstone form 30–40 m vertical cliffs prone to rock falls and slippage, although now largely protected from basal erosion by highway and railroad construction. From San Mateo Creek for 18 km southward to Aliso Canyon, seacliffs are formed in incompetent Miocene sands and gravels overlain by Pleistocene terrace deposits. These materials yield dissected landslide terrain in the 30–45 m San Onofre Bluffs fronting Camp Pendleton. Dissected terrace deposits continue to form low bluffs for 22 km from Aliso Canyon to Batiquitos Lagoon, the erosion of which has necessitated extensive use of riprap and seawalls to protect unwise bluff-top development (Fig. 1).

From Batiquitos Lagoon for 23 km southward to La Jolla, cliffs are developed mostly in Eocene marine and lagoonal sediments, ranging from conglomerates and coarse sandstones to siltstones and shales, capped by Pleistocene terrace deposits including iron-cemented paleodunes. These cliffs are commonly 20–30 m high but in the Torrey Pines area resistant sandstones reinforce 7 km of cliffs up to 100 m high. Here, as rock strength is weakened by percolating groundwater, large rock falls and landslides occur. At Blacks Beach, for example, a segment of cliff 140 m long, 60 m high and 50–80 m deep failed sometime between 1917 and 1922, and a further 230-m long mass involving 1.4×10^6 m³ of material collapsed onto the beach in 1982 (Vanderhurst et al. 1982). At Solana Beach farther north, cliffs developed in Pleistocene barrier-lagoon sediments suffer cavernous weathering from the solution of carbonate cements in former beach and dune sands (Kuhn and Shepard 1984).

At La Jolla and around Point Loma, relatively resistant Cretaceous sandstones, with subsidiary shales and conglomerates, support near-vertical seacliffs 10–30 m high, backed on Point Loma by Pleistocene seacliffs rising to over 100 m. At Sunset Cliffs these rocks dip 3–10° east and are broken by intersecting vertical joint sets which favor surge channels and sea caves extending up to 20 m inland (Kennedy 1973). These features help to explain why 5% of Sunset Cliffs retreated up to 3 m between 1896 and 1971, whereas 20% saw limited erosion and 75% no significant change. Erosion there increased slightly after construction of the Mission Bay entrance jetties in 1951 reduced littoral drift, but this was countered by revetments and cave closures completed in 1971.

Temporal Variability

The temporal variability of seacliff retreat largely reflects the frequency of storminess along the coast. Winter storms characterized by high seas and heavy rains may cause significant retreat from both mass movement and gulying of the cliff face and basal marine erosion, while persistent high water tables may induce subsequent mass movement at depth. Thus the wet and stormy period 1884–1893, which some have linked to the 1883 eruption of the volcano Krakatoa in the East Indies, led to extensive cliff retreat along the northern San Diego County coast. At Encinitas, several bluff-top city blocks were progressively devalued and had been removed from the tax assessor's records by 1895 (Fulton 1981). At Del Mar, bluff collapse seriously damaged the Southern California Railroad (now Atcheson, Topeka and Santa Fe Railroad) that had been completed in 1885 (Kuhn and Shepard 1980).

Similar wet and stormy years in the early 1920s and the period 1938–1947 also caused extensive cliff retreat. In late December 1940, a succession of storms accompanied by heavy rains, 10 m seas and high spring tides caused widespread damage from Redondo Beach to San Diego. Bluff collapse at Del Mar toppled a Santa Fe freight train towards the beach, mass movement talus was removed and cliffs eroded at Torrey Pines, and at Encinitas a recently built temple fell onto the beach (Kuhn and Shepard 1979). In June 1941, a segment of San Onofre Bluffs 500 m long slipped into the surf.

During the period 1948–1977, however, winter storminess was less common along the coast. Broad beaches accumulated below many cliffs, cliff erosion was minimal, and vegetation grew to inhibit shallow slope failures. This relatively passive period also saw extensive bluff-top development.

More recently, the period 1978–1983 saw a return to storminess and cliff retreat, including destruction of numerous bluff-top and cliff-base buildings constructed during the previous passive interlude. During February–March 1978 and February 1980, storm sequences approaching southern California from the west and southwest pushed 5–7 m waves against the coast. Onshore winds and low atmospheric pressure combined with high astronomical tides to generate superelevated storm surges which caused much beach and cliff erosion, especially between Encinitas and Carlsbad in San Diego County. These storms were accompanied by heavy rains throughout the area, causing much flooding and mass movement along the coast and inland. Between December 1982 and March 1983, storm seas, strong winds and high tides again promoted widespread coastal erosion, although rainfall was less concentrated and mass movement less common. This was a period of

Table 1. Seacliff retreat rates in southern California.

Cliff materials*	Location	Time period	Mean retreat rate (m/yr)	Source
San Diego County				
Quaternary sediments	La Jolla	1918-1930	0.25-0.50	1
	La Jolla	1912-1975	0.09-0.33	2
Eocene sandstone and mudstone	Encinitas	1884-1980	0.11	3
Cretaceous sandstone and siltstone	La Jolla	1940-1979	0.01-0.20	2
	La Jolla	1912-1975	0.09-0.26	4
	Sunset Cliffs	1896-1971	0.01-0.04	5
Santa Barbara County				
Pliocene siltstone	More Mesa	1927-1947	0.25	6
Mio-Pliocene shale	Isla Vista	1967-1973	0.05-0.39	7
	Goleta Point	1923-1967	0.08	6
Miocene shale	Santa Barbara Mesa	1951-1965	0.20	6

* Materials in lower cliff; upper cliff is often developed in Quaternary deposits. Sources: 1 Vaughan 1932, 2 Emery and Kuhn 1980, 3 Hannan pers. comm., 4 Hannan 1975, 5 Kennedy 1973, 6 Norris 1968, 7 Cottonaro 1975.

relatively high sea levels along the California coast, linked to the El Niño effect, and resulted in 5-7 m of cliff retreat at several locations in northern San Diego County. A small intense storm that passed across southern California on January 16-18, 1988, during breakdown of the next El Niño effect, produced significant wave heights of 10 m with a recurrence interval of more than 200 years (Flick and Badan-Dangon 1989). This storm seriously damaged artificial structures, notably at Redondo Beach, but caused little cliff erosion, probably because maximum storm surge coincided with lower tide and rainfall, though intense, was localized and brief.

Table 1 presents data on seacliff retreat for several localities. Although harder rocks such as Cretaceous sandstone probably retreat more slowly than softer rocks, generalizations are difficult because of the temporal variability involved. Longer-term rates are a better measure of secular coastal behavior, integrating as they do the effects of storms and quiet periods. Measurements relating to a single storm may have little long-term significance. On the other hand, the episodic nature of cliff retreat must be understood for coastal management purposes. Bluff retreat at Solana Beach, for example, amounted to 3 m between 1971 and 1978, but this occurred in two brief episodes of 1.5 m (Kuhn and Shepard 1979, 1984). Large quantities of talus were eroded from the cliff base at Blacks Beach in December 1940, followed by new debris accumulations which were not wholly removed until winter 1978.

Longer-term rates of seacliff retreat along the San Diego coast range from 0.01 to 0.26 m/yr in compact Cretaceous and Eocene sandstone and siltstone to 0.09 to 0.50 m/yr in Quaternary alluvium. In localities protected by a broad beach, however, soft sediments may suffer little erosion, where in exposed areas hard rocks may retreat quite rapidly along structural weaknesses until the bounding rocks collapse into sea caves and surge channels (Kennedy 1973).

On the south-facing coast between Santa Barbara and Isla Vista, 10-60 m high

seacliffs are mostly sheltered from storm seas but weak lithologies and surface runoff still favor cliff retreat. Between 1951 and 1965, friable Miocene shales fronting Santa Barbara Mesa retreated at a mean rate of 0.20 m/yr but 50 m cliffs in more siliceous beds to the west had showed little change since 1927 (Norris 1968). Soft fractured Mio-Pliocene shales retreated 0.08 m/yr at Goleta Point between 1923 and 1967 (Norris 1968), but to the west at Isla Vista retreat ranged from 0.05 to 0.39 m/yr from 1967 to 1973 (Cottonaro 1975). Pliocene siltstone and shales fronting More Mesa retreated 0.25 m/yr between 1927 and 1947.

Rates of seacliff retreat have been modified in recent decades by human attempts to protect property, roads and railroads. Emplacement of riprap and seawalls at the cliff base, concrete crib structures against cliff faces, groin construction along fronting beaches, and sealing of sea caves have all been employed. Some 20 km of the San Diego County coastline are now protected one way or another. Such structures have had mixed success, often deflecting wave energy to nearby unprotected cliffs while having little impact on mass movement and gulying.

Conclusions

Seacliff retreat represents a complex response between the processes directed at the land-sea interface and the properties of the coastal terrain. With so much variety in the location and timing of this response, it is difficult to translate a body of empirical knowledge into a universal model that will predict exactly where and how fast seacliffs will retreat. Nevertheless, significant advances have been made in recent years concerning recognition of the problems and the range of solutions necessary for effective coastal zone management.

With respect to the problems, the episodic nature of seacliff retreat is clearly related to meteorological conditions, especially the coincidence of storm seas and heavy persistent rainfall as in 1978 and 1980. Without exceptional rainfall, storm seas will still cause marine erosion at the cliff base, especially during high tides and El Niño events, but mass movement is less widespread. On the other hand, deep-seated landslides may occur several months after heavy rains, or be triggered independently by excessive water from septic tanks, road drainage, or landscape irrigation. Earthquakes may cause cliff collapse at any time. Sea-level rise or increased storminess associated with climatic change may raise the frequency of seacliff retreat but, in the short term, these influences are small compared with the pervasive impact of human interference. As stated earlier, human activities may often recreate the groundwater hydrologies and stress fields of wetter Pleistocene climates, pushing coastal bluffs of marginal stability over the threshold toward renewed failure.

At the coast, as elsewhere, it is the role of earth scientists to recognize the relationship between geomorphic processes and the properties of underlying materials, to identify unstable and potentially unstable terrain, and to advise individuals and government agencies regarding these hazards. It is the duty of agencies responsible for land-use decisions to seek such advice and act accordingly.

Much early development of the southern California coast proceeded in ignorance of the physical properties of coastal terrain, compounded by a false sense of security created during three decades of relatively dry conditions following World War II. Development was permitted dangerously close to unstable cliffs and on potentially unstable coastal slopes. Reactivation of the Portuguese Bend

Landslide in 1956 and subsequent findings of government liability did, however, eventually lead various cities and counties to enact ordinances that required geologic reports in areas of perceived hazard. Even then, the variable nature of these reports and the uneven application of the ordinances initially did little to inhibit unwise development. Following the citizen initiative of Proposition 20 in 1972, the California Coastal Act of 1976 provided a more rational framework for coastal land-use decisions. The Beverly Act of 1979 enabled creation of Geologic Hazard Abatement Districts such as those formed to address the Abalone Cove and Klondike Canyon landslides in 1981 and 1982 respectively. In recent years, however, diminished state support for coastal management has combined with variable local government commitment and changing federal priorities to weaken the California Coastal Act. Furthermore, although the act recognized the problem of seacliff retreat, scientists and engineers continue to differ regarding the best solutions for mitigating slope failure and cliff collapse. The act also intensified rather than resolved the conflict between public use of the coast and private property rights, creating a difficult climate for coastal management.

Seacliff retreat is a natural process which, if unheeded, threatens human life and livelihood, and which can be aggravated by human activity. It will continue to occur and therefore responsible coastal management must require that human activity be set back an appropriate distance from cliff tops and diverted from unstable and potentially unstable terrain. Above all, whereas we cannot presently influence the magnitude and frequency of storm rainfall, much can be done to minimize concentrations of runoff and other waters on seacliffs and coastal slopes.

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Research Notes

The Gulf Coney, *Epinephelus acanthistius*, from the Marine Waters of Southern California

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The gulf coney, *Epinephelus acanthistius* (Gilbert, 1892) (Fig. 1), was originally described from a single specimen collected by the ALBATROSS along the eastern shore of the Gulf of California. For a number of years it was considered a rare species and was listed as such by Walford (1937), "The only specimen known is one 16 inches long, found in the Gulf of California in 1890." Walford (1936, 1937) recognized two species of coney, the gulf coney (*Cephalopholis acanthistius*) and one which he described as *Cephalopholis popino*, rose coney, with a type-locality of Mazatlan, Mexico. Subsequently, Smith (1971) synonymized these nominal species, stating "The characters by which Walford distinguished *Cephalopholis popino* from *C. acanthistia* are variable in other genera and I see no reason to suppose they are diagnostic here." This action was followed by Walford (1974) in his revised edition of "Marine Game Fishes." Within the last 20 years, gulf coney has become an important food fish and large quantities are imported into California and sold primarily to restaurants under the Mexican name of baqueta. Imports of gulf coney between 1984 and 1989 ranged from 136,393 to 897,999 pounds (Patricia J. Donley, Nat. Mar. Fish. Serv., pers. comm.). It is an extremely fine flavored fish and commands a high market price (David Ptak, Chesapeake Fish Company, San Diego, pers. comm.). The range of the species has been recorded as the Gulf of California to Peru (Thomson et al. 1979), a typical Panamic distribution. Thomson et al. (loc. cit.) also list it as "a common bottom fish throughout the Gulf of California." Two captures of this species from off southern California between December 1988 and November 1989 allow us to document the occurrence of the gulf coney from off our state.

On 17 December 1988, while fishing for bottom fishes off Whites Point, Los Angeles County, the junior author landed an unusual fish which was taken to Cabrillo Marine Museum where it was tentatively identified as a gulf coney. The fish measured ca. 46 cm total length (TL), and was maintained alive for public display until it died in June 1990. On 28 November 1989 a second gulf coney was captured on hook-and-line by Zora Weinhardt while fishing off the barge ISLE OF REDONDO in Santa Monica Bay. The ISLE OF REDONDO is anchored on the edge of Redondo Canyon in ca. 60 m of water. This specimen, measuring 495 mm TL and weighing 2.30 kg, is deposited in the Section of Fishes, Natural History Museum of Los Angeles County, LACM 44983-1. Counts and measurements in mm (percent of standard length in parentheses) are as follow: D IX, 17;

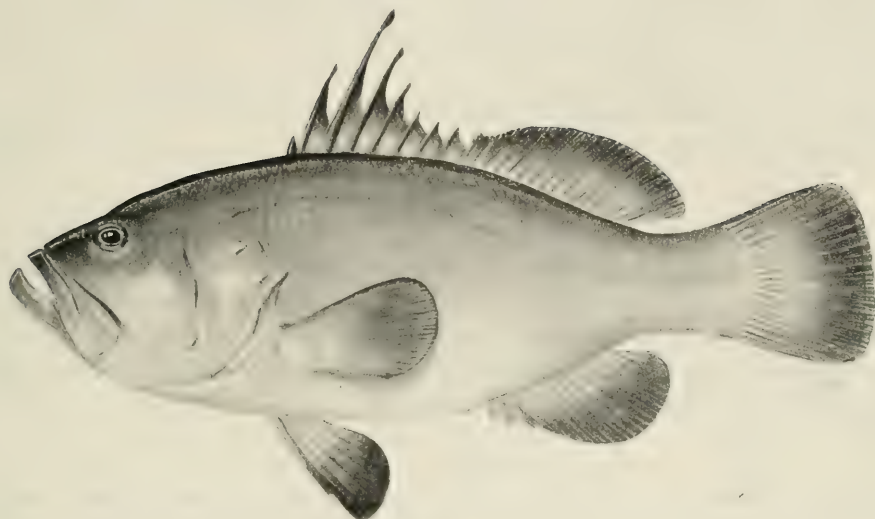


Fig. 1. Illustration of gulf coney, 710 mm total length. Taken from Reef Fishes of the Sea of Cortez; Thomson et al., 1979. By permission of the artist, Alex N. Kerstitch.

A III,9; pect. 18/18; gill rakers $9+23=32$ (including rudimentary rakers); standard length 393 mm; head length 166.7 (42.4); snout length 42.4 (10.8); orbit width 28.4 (7.2); bony interorbital width 24.0 (6.1); suborbital height 18.6 (4.7); pectoral length 100.1 (25.5); pelvic length 98.4 (25.0); pelvic spine length 52.9 (13.5); first dorsal spine 18.5 (4.7); second dorsal spine 56.3 (14.3); third dorsal spine 86.2 (21.9); fourth dorsal spine 78.8 (20.1); fifth dorsal spine 64.0 (16.3); second anal spine 30.3 (7.7); third anal spine 45.5 (11.6); body depth at origin of pelvic fin 159.5 (40.6); base of dorsal fin 205.0 (52.2); base of anal fin 76.0 (19.3); base of pectoral fin 28.1 (7.2); caudal peduncle depth 56.0 (14.2); dorsal length of caudal peduncle 56.0 (14.2); ventral length of caudal peduncle 71.0 (18.1); length of gill raker at angle 19.0 (4.8). In addition, among eastern Pacific *Epinephelus*, the plain rose-colored body, elongate anterior dorsal spines (II–V), and moustache just above the maxilla are diagnostic characters.

The capture of gulf coney, as well as three other species of Panamic sea basses (family Serranidae—representing the genera *Epinephelus*, *Paranthias*, and *Serranus*) off southern California (M. James Allen, Robert J. Lavenberg, and Richard H. Rosenblatt pers. comms.) raises the question as to the occurrence of these fishes off this area, biogeographically a warm-temperate environment. Whether these fishes were transported into our waters during the El Niño event of 1982–84 and are now (6 to 7 years later) just entering sportfish catches can only be a matter of speculation. Expatriation during a warm-water event seems a reasonable explanation for a stone scorpionfish (*Scorpaena mystes*) collected off Los Angeles County in 1984 (Swift 1986) and Cortez angelfish (*Pomacanthus zonipectus*) which

were first noted from Californian waters in 1985 (Lea et al. 1989). In the case of gulf coney, and the other sea basses noted above, other phenomena may be responsible. There has been much interest of late in the concept of global warming (Houghton and Woodwell 1989) and the subsequent increase in ocean temperatures. Changes in ocean temperatures would be expected in turn to be followed by changes in distributions in the marine flora and fauna. At this point the mechanism or mechanisms responsible for the recent arrival of Panamic elements off California remains conjectural.

Acknowledgments

We would like to thank Stephen J. Crooke, California Department of Fish and Game; Lloyd Ellis, Cabrillo Marine Museum; Richard H. Rosenblatt, Scripps Institution of Oceanography; Camm C. Swift and Robert J. Lavenberg, Natural History Museum of Los Angeles County; and M. James Allen, Marine Biological Consultants; for their constructive comments and information provided during the preparation of this research note.

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The Ontogenic Acquisition of Infestation of the Trematode Ectoparasite *Neobenedenia girellae* on the Marine Teleost *Girella nigricans*

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The opaleye, *Girella nigricans*, is a kyphosid fish that ranges from San Francisco, California, to Cape San Lucas, Baja California, inhabiting shallow reefs and kelp beds to depths of 29 m (Gotshall 1981). Infestation of *Girella nigricans* by the monogenetic trematode, *Neobenedenia girellae* (Hargis), was observed previously off La Jolla, California (Hargis 1953). Thirty-nine *N. girellae* were isolated from an unspecified number of *G. nigricans* in the original description of this trematode species of the family Capsalidae (Hargis 1953). This ectoparasite has been found to infest the staghorn sculpin (*Leptocottus armatus*), California sheephead (*Semicossyphus pulcher*), and the grouper (*Mycteroperca pardalis*) in addition to *Girella nigricans* (Yamaguti 1963).

Little is known about the biology of *N. girellae*. This study was undertaken to investigate the temporal aspects of the interaction between host and parasite and to quantify the degree of infestation of *G. nigricans*.

Ten juvenile *G. nigricans* (0+ years old; 40.0–53.0 mm standard length (SL)) and ten subadults (1–2+; 78.6–107.1 mm SL) were obtained in Big Fisherman's Cove, Catalina Island, CA using a 25 mm mesh seine. A 38 mm mesh gill net was set in the same cove to collect 14 younger adults (2+–4+; 114.0–128.8 mm SL). Eleven older adults (3+–7+; 170.6–270.6 mm SL) were taken by spear and gill net (75 mm mesh) near Bird Rock, Catalina Island, CA. All collections were made during October, 1988.

Live specimens were placed in holding tanks at the Catalina Marine Science Center laboratory. The speared adults were refrigerated until inspection (no more than eight hours after capture). The live fish were inspected within two days after capture.

Each fish was measured (standard length) and its age determined by counting scale annuli (Cailliet et al. 1986). The body and fins were inspected visually for external trematodes. These were placed in separate vials of AFA (10 parts formalin, 50 parts 95% ethanol, 2 parts glacial acetic acid, and 40 parts distilled water, following Cailliet et al. 1986), and were later transferred to 70% ethanol. Trematodes were identified following Crane (1972). Voucher specimens of *N. girellae* were deposited in the collection of the Harold W. Manter Laboratory, University of Nebraska State Museum (Nos. HWML 32720–32722).

The four nonoverlapping size categories; juveniles, subadults, young adults, and old adults, contained a minor overlap in age as ascertained by scale annuli. A standard linear regression between size and age of *G. nigricans* was calculated (Fig. 1) and showed a high degree of correlation ($r = 0.949$, $P < 0.001$). There appears

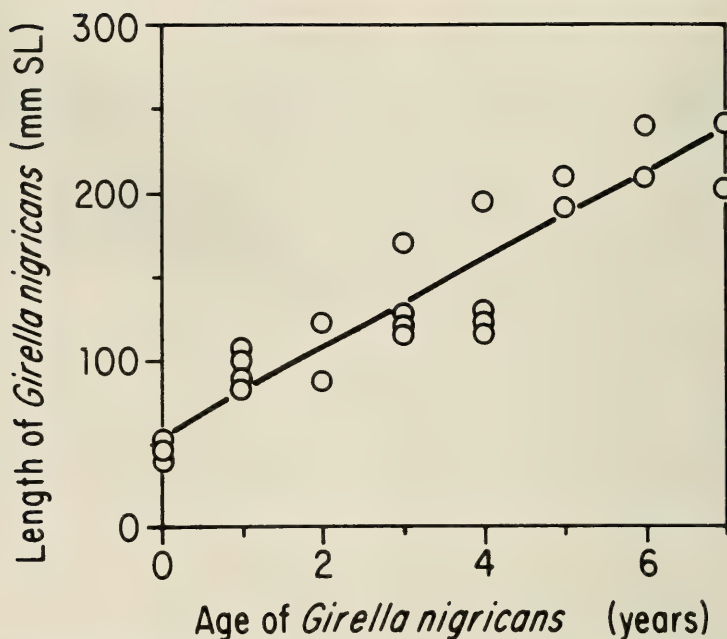


Fig. 1. Simple linear regression correlating size (standard length) with age of *Girella nigricans* ($r = 0.949$, $P < 0.001$).

to be a minimum age when fish become parasitized. This occurs at approximately 200 mm SL (a minimum age of 5 yrs). *Girella nigricans* below this threshold are not parasitized. In the older adult size class, 7 of 8 specimens (87.5%) of five years of age or older were parasitized (with 2, 4, 6, 8, 10, 15, and 18 *N. girellae*). A significant correlation was shown between the size of *G. nigricans* and number of parasites for the parasitized specimens ($t = 23.48$, $df = 12$, $P < 0.001$). Too few specimens were analyzed to determine a correlation between sex and infestation.

How might we account for this size variability in parasite infection? First, with time fishes probably encounter more parasites. Second, larger fishes have a larger surface area with which to accommodate higher degrees of infestation (Cailliet et al. 1986). Third, physiological differences between juveniles and adults, which may occur at or after age 5+, may prevent attachment of the parasite or lead to death of the younger individual if attachment does occur. If the latter were true, all parasitized fish younger than 5+ years old might die as a result of infestation. This would account for the absence of parasitized fish in these age categories.

It is known that juvenile and subadult *G. nigricans* seek out increasingly deeper water as they mature (Feder et al. 1974), and it is possible that differences in the habitats of juvenile and adult *G. nigricans* may account for the absence of parasites on the juveniles. However, 1 of the 8 adults (age 5 and older) examined was not parasitized as were the younger adults (age 3+–4+) collected in the same area. Thus, it is unlikely that habitat plays a major role and supports the idea that infestation may be related to age. As the fish mature, their chances of encountering trematodes increase, leading to increased infestation. Environmental parameters may be another factor. The older adults may have grown up during a period when external conditions (i.e., temperature, salinity, turbidity, etc.) favored the acqui-

sition of parasites. These conditions may have changed, preventing reacquisition of the trematodes by the younger fish. Thus, only older fish, which grew up in favorable conditions, would be parasitized. Additional data must be collected to test whether infestation is specific to age-class (5+ yrs old) or year-class (1983 and earlier).

It is obvious that there is a lot more to learn from this host-parasite interaction. More host specimens over 200 mm SL should be examined to determine how these trematodes are acquired and to better quantify the degree of infestation of this vulnerable class size.

Acknowledgments

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A New Sepiamorph Sepiid (Mollusca: Cephalopoda) from the Eocene of California

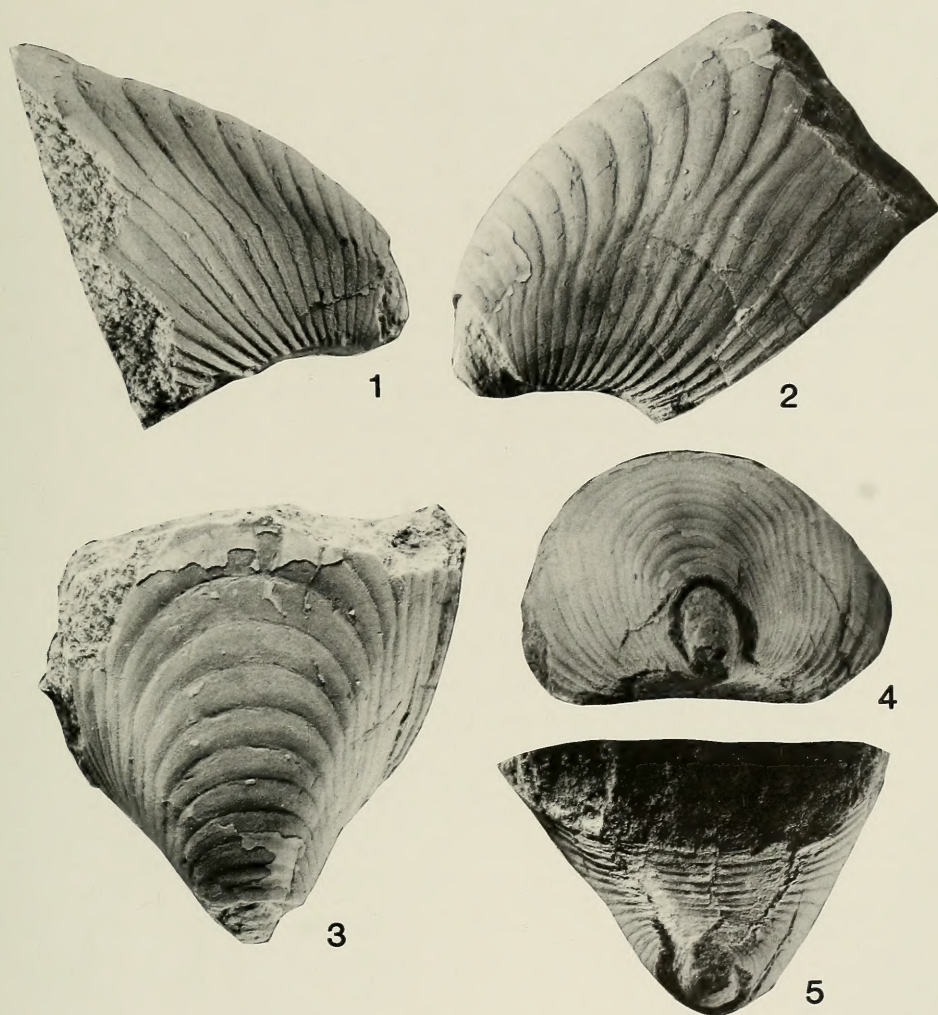
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Sepiids are extremely rare in the Paleogene fossil record of the Pacific coast of North America, and only three specimens have been previously reported. One is a partial phragmocone of an unidentifiable spirulimorph sepiid from the "Stewart bed" in the lower middle part of the Lajas Formation, Ventura County, southern California (Squires 1983, figs. 2J–K; 1984, fig. 13b). The other two are partial phragmocones of an unidentifiable sepiamorph sepiid from the upper Eocene Hoko River Formation, Olympic Peninsula, northwestern Washington (Squires 1988, figs. 2.3, 3.9–3.14).

Recently, G. L. Kennedy of the Natural History Museum of Los Angeles County Invertebrate Paleontology Section (LACMIP) brought to my attention that the museum has a sepiid specimen from Eocene strata in Rose Canyon just southeast of the intersection of Interstate Highway 5 and highway 52 in the northern part of San Diego, San Diego County, southern California. No precise locality information is known for this specimen, but the only Eocene strata that crop out in Rose Canyon are the middle Eocene Ardath Shale and the overlying middle Eocene Scripps Formation (Givens and Kennedy 1979, figs. 1, 3, 6). The grayish green rock matrix surrounding the very slightly crushed specimen is a well indurated, very fine-grained sandstone that is similar to rocks found in both the Ardath Shale and the Scripps Formation.

The Rose Canyon sepiid specimen is an internal mold with remnants of a thin nacreous layer on one side (Figs. 1–5). The specimen, which is missing the rostrum and the living chamber, consists of a partial phragmocone 22 mm in length with a maximum height of 18 mm. The apical angle is 60°. At a distance of about one-third the length of the phragmocone from the apical region, there is a slight constriction of the lateral walls of the phragmocone. The dorsal surface of the phragmocone is broadly rounded, but the ventral surface is only half as long and, although flattened relative to the overall phragmocone, is distinctly arched. There are 17 closely spaced chambers that are very narrow near the ventral surface but much wider dorsally. The height to maximum diameter ratio of the chambers is approximately 1.36, and the maximum diameter of each chamber increases in size by 11 percent relative to the preceding younger chamber. The septa of the chambers meet the ventral surface at an angle of about 80° near the apical region, but the angle decreases to about 50° at the oral end of the ventral surface. The septal suture has a very slight ventral lobe and a distinct lateral lobe that raises addorsally into a broad and rounded dorsal saddle (Fig. 6). The siphuncle, which is broken and poorly exposed in the apical region, is large and seems to have an endogastric curvature and rapidly expands adorally. It is on the ventral side of



Figs. 1-5. Sepiamorph sepiid, hypotype, LACMIP 8407, middle Eocene strata, Rose Canyon, San Diego County, southern California, $\times 2.8$ unless otherwise noted; 1, left lateral view; 2, right lateral view; 3, dorsal view; 4, apical view, $\times 2.4$; 5, ventral view, $\times 2.4$.

the phragmocone. The cross-sectional shape of the phragmocone is elliptical and hoof-like.

The Rose Canyon sepiid belongs to *Sepia*-like sepiids with strongly oblique (slanting) sutures. These sepiamorphs include *Belosepia*, *Sepia* (Jeletzky 1969, p. 107) and the Hoko River sepiid (Squires 1988). The Rose Canyon sepiamorph most closely resembles *Belosepia* Voltz, 1830, even more than does the Hoko River sepiid. The ventral flattening of the phragmocone, the hoof-like cross section, the lateral lobe, and the rapidly expanding siphuncle of the Rose Canyon specimen also are present in *Belosepia*, and these features help characterize this genus (Edwards 1849, Pl. 1, figs. 1h, 5, 6; Jeletzky 1969, pp. 26-27). The Rose Canyon specimen differs from *Belosepia*, however, in having only a very slight

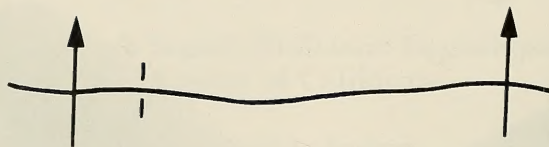


Fig. 6. Septal suture of the Rose Canyon sepiamorph sepiid, hypotype, LACMIP 8407, at chamber height of 15 mm along left-lateral side of phragmocone, $\times 2.8$. The septal suture was drawn from photographs.

ventral lobe. *Belosepia* normally has a well developed ventral lobe (Jeletzky 1969, p. 26). The Rose Canyon sepiid is at least specifically distinct and possibly generically distinct, but it is doubtful that one should erect any new taxa for the Rose Canyon specimen because of the fragmentary and otherwise poor preservation.

Jeletzky (1969) mentioned that *Sepia* (family Sepiidae Keferstein) separated from *Belosepia* (family Belosepiidae Naef) during the middle Eocene. The Rose Canyon sepiid probably belongs in family Belosepiidae, but is nearer the separation of this family into Sepiidae, than any previously known sepiid.

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COVER: Gulf Coney, *Epinephelus acanthistius*, taken from Reef Fishes of the Sea of Cortez; Thompson et al., 1971. By permission of the artist, Alex N. Kerstitch. (See page 80)